

Climate-carbon feedback of the high latitude ocean



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Abstract

The warming induced by anthropogenic carbon emissions affects the climate system through a multitude of physical mechanisms. Changes in the dynamics, thermodynamics and biogeochemistry of the ocean alter the different ocean carbon reservoirs, potentially resulting in further carbon emissions and a climate-carbon feedback. Surface wind stress and surface warming are two of the most influential forcings acting on the ocean carbon system in past, present and future climates due to their influence on the mixed layer dynamics and the large scale ocean circulation.

This thesis quantifies the climate-carbon feedback of wind stress and surface warming, with a particular focus on the mechanisms driving the feedbacks and the role of the Southern Ocean and the North Atlantic. To study the feedbacks, a set of theoretical scalings and a hierarchy of numerical simulations are used. Of the climate feedbacks examined, increased surface warming is likely to result in large atmospheric CO₂ anomalies while the effects of North Atlantic wind stress are likely to be negligible. The atmospheric feedback of surface warming is constrained by compensating changes in separate ocean carbon reservoirs as a result of warming-induced circulation changes. Southern Ocean winds affect atmospheric CO₂ through both local upwelling of carbon as well as the remote modification of Equatorial and North Atlantic chemistry. As a result, the net Southern Ocean wind stress feedback could be significant and even comparable to the temperature feedback. This thesis provides a quantification of regional and global climate-carbon feedbacks due to ocean dynamics. The estimates of carbon-climate feedbacks are useful tools for understanding past, present and future climates.

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List of Abbreviations

Abbreviation	
AABW	Antarctic Bottom Water
AAIW	Antarctic Intermediate Water
ACC	Antarctic Circumpolar Current
ALK	Alkalinity
AOU	Apparent Oxygen Utilization
CMIP3/CMIP5	Coupled Model Inter-comparison Project phase 3/5
CanESM2	Canadian Earth System Model 2
DIC	Dissolved inorganic carbon
DOP	Dissolved organic phosphorous
EKE	Eddy kinetic energy
GCM	General Circulation Model
GFDL-ESM2M	Geophysical Fluid Dynamics Laboratory Earth System Model 2
HadGEM2-ES	Hadley Centre Global Environment Model 2, Earth System version
IPCC	Inter-governmental Panel for Climate Change
IPSL-CM5	Institut Pierre Simon Laplace Climate Model 5
MOC	Meridional Overturning Circulation, the upper overturning cell in the Atlantic
NA	North Atlantic sub-polar gyre
NADW	North Atlantic Deep Water
NAM	Northern Annular Mode
NOAA	National Oceanic and Atmospheric Administration
NorESM	Norwegian Earth System Model
OMZ	Oxygen Minimum Zone
ppm	parts per million
SAM	Southern Annular Mode
SO	Southern Ocean
SST	Sea surface temperature
Sv	Sverdrup, $1 \text{ Sv} = 10^6 \text{ m}^3/\text{s}$

List of Recurring Symbols

Symbol	
$[\dots]_h$	$\int_{z=-H}^{z=\eta} \dots dz$
$\langle \dots \rangle$	Temporal mean of the control run
$(\)^n$	Normalized anomaly, where the spatial anomaly is divided by the global average of the anomaly
$(\)'$	Anomaly with respect to the control or pre-industrial state
∇	2D vector operator
∇	3D vector operator
$+T_N$	Northern boundary warming experiment, such that T_N is multiplied by a factor $4\beta - 3$
$++T_N$	Northern boundary warming experiment, such that T_N is multiplied by a factor $6\beta - 5$
$+T_S$	Southern boundary warming experiment, such that $T_S = 2(\beta - 1)$
$2\times\text{CO}_2$	CMIP5 simulation whereby atmospheric CO_2 is instantaneously doubled and held constant
$4\times\text{CO}_2$	CMIP5 simulation whereby atmospheric CO_2 is instantaneously quadrupled and held constant
β	The magnitude of the global temperature perturbation, so a $1.2\times T$ experiment has $\beta = 1.2$
Δ	Anomaly with respect to the mean state
$\Delta_x, \Delta_y, \Delta_z$	Spatial difference along the zonal, meridional or vertical direction
$\Delta_z T$	Temperature difference between the mixed layer and underlying water ($^{\circ}\text{C}$)
ϵ_f	Non-local feedback parameter
$\mathbf{f}$	Forcing applied to ocean's surface
γ	Nutrient limitation factor: determines the strength of surface biological productivity based on nutrient and light limitation, varies between 0 and 1
κ_{GM}	Isopycnal eddy diffusivity in Griffies [1998] formulation of Gent and McWilliams [1990] eddy parametrization
ϕ	Proportionality between C_N and C_S in preformed carbon transport box model, depends on f and Ψ
$\overline{\Psi}, \Psi^*, \Psi$	Eulerian, Bolus and Residual overturning streamfunction, $\Psi = \overline{\Psi} + \Psi^*$
σ	Potential density
τ	Surface wind stress
τ_{SO}	Surface wind stress in the Southern Ocean
τ_{NA}	Surface wind stress in the North Atlantic sub-polar gyre
Alk_{pre}	Preformed Alkalinity
Alk_{reg}	Regenerated Alkalinity
C	Dissolved inorganic carbon concentration
C_{ant}	Anthropogenic carbon
C_{carb}	Calcium carbonate
C_d	Drag coefficient
C_D	Deep ocean carbon concentration in preformed carbon transport box model

C_{dis}	Disequilibrium carbon
C_h	Mixed layer dissolved inorganic carbon concentration
C_N	(Northern) surface ocean carbon concentration in preformed carbon transport box model
C_p	Specific heat capacity of sea water at constant pressure (4000 J/kg/K)
C_{pre}	Preformed carbon
C_S	(Southern) surface ocean carbon concentration in preformed carbon transport box model
C_{sat}	Saturated carbon
C_{soft}	Soft-tissue carbon
C_{reg}	Regenerated carbon
e_{GM}	Proportionality parameter between meridional Ekman and bolus velocities in the mixed layer
f	Strength of vertical mixing in the ocean
$F_{air-sea}$	Air-sea flux of CO ₂ (positive into ocean)
F_{bio}	Biological flux of CO ₂
g	Gravitational acceleration (9.8 m/s ²)
h	Mixed layer depth
I_B	Buffered carbon inventory
k	Piston velocity in Wanninkhof [1992] parametrisation for the air-sea flux of CO ₂
m_0	Constant coefficient measuring sea surface roughness
N	Total nutrient concentration
N^*	Efficiency of the biological pump, $N^* = \frac{N_b}{N}$
N_b	Regenerated nutrient concentration
N_p	Preformed nutrient concentration
O_2	Oxygen
pCO_2^a	Partial pressure of CO ₂ in the atmosphere (well mixed)
pCO_2^o	Partial pressure of CO ₂ at the oceans surface
P_E	Wind stress-weighted area fraction of rest of ocean where a forcing f is not applied : $P_E = \frac{A_E \overline{\tau_E}}{A_H \overline{\tau_H} + A_E \overline{\tau_E}}$
P_H	Wind stress-weighted area fraction of region where a forcing f is applied : $P_H = \frac{A_H \overline{\tau_H}}{A_H \overline{\tau_H} + A_E \overline{\tau_E}}$
R	Revelle Buffer Factor
S_N	Northern boundary restoring salinity
S_P	Nutrient stream: $S_P = [PO_4^{3-}] v$
S_{Px}	Zonal nutrient transport: $S_{Px} = [PO_4^{3-}] u$
S_S	Southern boundary restoring salinity
T_G	Global mean sea surface temperature
T_N	Northern boundary restoring temperature
T_S	Southern boundary restoring temperature
$\bar{\mathbf{u}}, \mathbf{u}^*, \mathbf{u}$	3D Eulerian, bolus and residual velocities
u_*	Frictional zonal wind velocity $u_* = \left \frac{\tau}{\rho_0} \right ^{1/2}$ (m/s)

Introduction

Since the start of the industrial era, atmospheric carbon dioxide concentrations have risen from 280 ppm to breach 400 ppm in 2013, representing an increase of over 40%. Carbon dioxide, CO_2 , is an important greenhouse gas so any increase in global concentrations has a substantial impact on global temperatures. The increase in the atmospheric CO_2 reservoir only represents a part of the emitted anthropogenic carbon, since most of the carbon is taken up and stored by the ocean. The ocean is by far the largest active reservoir of carbon at 40,000 Gigatonnes (GtC), while the atmosphere contains only 750 GtC and the biosphere just over 2000 GtC. On timescales of several centuries, around 80% of emitted anthropogenic carbon will be mixed into the ocean [Solomon et al., 2009].

Fully appreciating the effect of anthropogenic CO_2 on the climate requires understanding how ocean dynamics and the carbon cycle respond to climate change. Section 1.1 briefly discusses the key dynamical processes within the ocean that determine the internal distribution of carbon. Section 1.2 describes the different components of the ocean carbon cycle and how they are affected by these dynamics. Section 1.3 outlines previous work in quantifying dynamical feedbacks of the ocean carbon cycle on atmospheric CO_2 and section 1.4 discusses how this thesis aims to further our understanding of climate feedbacks.

1.1 Ocean processes controlling carbon storage

1.1.1 Mixed layer

The interaction between the ocean and atmosphere is dictated by the mixed layer, the surface layer in the ocean stirred by the surface winds such that water density and tracers are near-uniform within. The mixed layer's carbon concentration and temperature set the magnitude of the atmosphere-ocean exchange of carbon and heat at the ocean surface, respectively. The main fluxes in and out of the mixed layer that set the local properties are generally: air-sea surface fluxes, lateral and vertical advection, and lateral and vertical diffusion. The heat flux is dictated by air-sea temperature differences and for carbon by the air-sea disequilibrium in partial pressure of CO_2 , $p\text{CO}_2$. The lateral advection is

given predominantly by the wind driven surface circulation. The lateral advection, which depends on the background tracer gradient, can actively modify tracer concentrations in the mixed layer if surface currents are across sloping mixed layer boundaries.

Vertical advection in the mixed layer is often thought about in terms of entrainment due to changes in the mixed layer depth. Due to the seasonal changes in radiative and freshwater forcing, the mixed layer depth exhibits a seasonal cycle of shallowing and deepening. The two main drivers of mixed layer depth are density and wind stress. In the winter, surface cooling leads to densification of mixed layer water and stronger winds cause a deeper stirring, which deepens the mixed layer base. As a result, water underlying the summer mixed layer is entrained and mixed into the winter mixed layer. Deep mixed layers are a signature of a connection between the surface and deep ocean, exchanging water properties between both. The driving processes of mixed layer deepening act as a thermostat and stabilise mixed layer temperatures [e.g. Xie et al., 2010]. For example, wind induced mixing in the Southern Ocean leads to deep mixed layers and brings deep carbon-rich water to the surface. In regions of strong air-sea surface heat flux, e.g. the Tropics, low surface densities cause shallow mixed layer and isolation of the surface ocean from underlying waters.

Diffusive fluxes also need to be accounted for in the mixed layer, but are usually much smaller compared to the other terms [e.g. Levy et al., 2013]. There are also tracer-specific mixed layer fluxes that do not explicitly depend on the circulation such as a radiative heat loss in the case of temperature or a biological flux in the case of carbon or oxygen. These will be discussed later on.

1.1.2 Surface circulation

The ocean’s surface circulation is largely determined by the forcing of zonal surface winds, which act as a source of momentum. Ocean gyres are created through Sverdrup balance¹ [Sverdrup, 1947]: anti-cyclonic (clockwise in the Northern hemisphere) rotating gyres in the sub-Tropics and cyclonic gyres in the sub-polar regions (see Fig. 1.1a). Through Ekman transport [Ekman, 1905], there is net transport of water at right angles to the surface current such that water is aggregated in the interior of the sub-tropical gyres and

¹The balance arises from the conservation of total vorticity, whereby a change in relative vorticity due to wind stress is balanced by a change in planetary vorticity.

exported out of sub-polar gyre leading to convergence and divergence, respectively. The net result is a rise in sea surface height in sub-tropical gyres and a sea surface depression in sub-polar gyres. The surface gyres are in geostrophic balance such that time mean surface currents mostly follow sea level height contours (see Fig. 1.1).

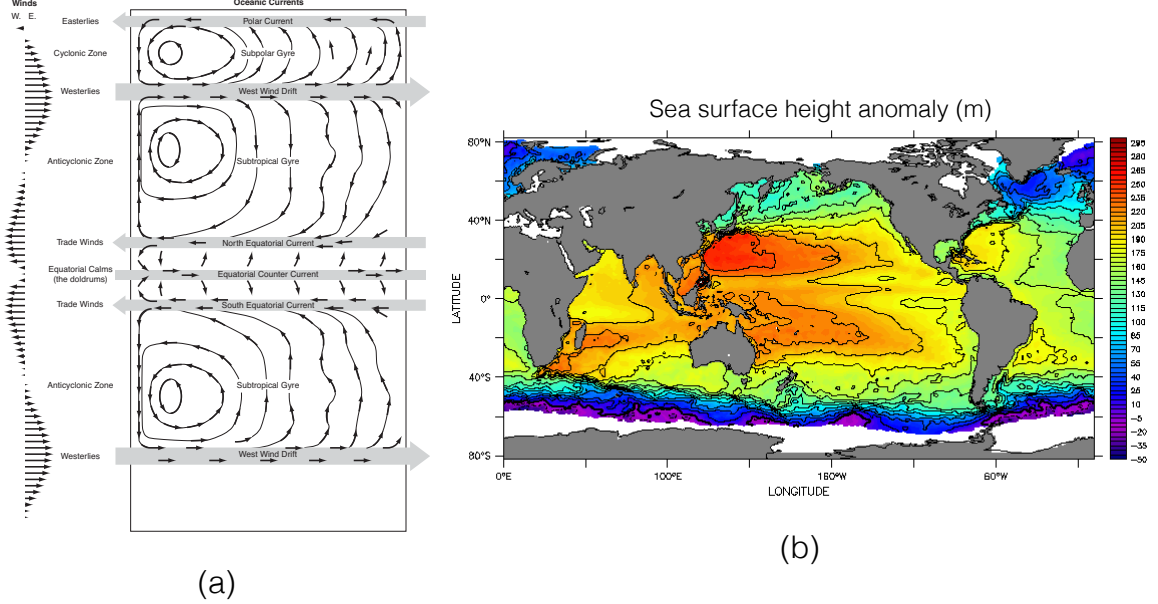


Figure 1.1: a) Theoretical gyre circulation diagram; the arrows indicate the flow pattern; b) Sea surface height anomalies relative to the global mean sea level for the period 1990-2000.

Convergence in sub-tropical gyres causes a local depression of the thermocline² causing mixing of surface tracers with deeper waters. Similarly, the divergence in sub-polar gyres raises the thermocline such that deep waters are entrained into the surface mixed layer causing upwelling of tracers. The thickness of the thermocline determines the heat and surface tracer content of the upper ocean, as well as the inertia of the interactions with the atmosphere on seasonal timescales. The thermocline depression in sub-Tropical gyres is visible in ocean tracers that are forced at the ocean's surface, such as salinity (Fig. 1.2a), where there is a depression of the signal in the mid-latitude sub-tropical gyres.

The vertical structure of the gyres in the stratified ocean also results in shoaling of vorticity contours in the interior of the gyre that end on continental boundaries rather than reaching the surface. As a result, there are regions in the eastern Tropics in each

²Generally defined as the depth of the maximum in the vertical gradient of temperature, signalling strong stratification.

ocean basin that are unventilated by the wind-driven circulation. These regions, generally between 200-1000m depth, are referred to as shadow zones. Due to the low ventilation (see Fig. 1.2b), shadow zones contain older water which accumulates organic tracers such as carbon and phosphate, and are subsequently stripped of oxygen. Shadow zones are therefore often measured in the real ocean through oxygen concentration and are referred to as oxygen minimum zones. Sverdrup balance in the sub-tropical gyre results in net equator-ward transport in the interior, balanced by a narrow western boundary current, where frictional forces are large enough to break the vorticity balance. Due to the high meridional transport and velocities, 30 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3/\text{s}$) at 1-2 m/s for the Gulf Stream approximately [Lund et al., 2006], there is strong heat flux out of the ocean (see Fig. 1.6c). Due to the water exchanged between the sub-polar and sub-tropical gyres, the western boundary currents and their extensions are vital in the meridional transport of ocean heat and tracers.

In the Southern high latitudes, the Southern Ocean forms an open channel around Antarctica. The westerly winds over the Southern Ocean accelerate a zonal Antarctic Circumpolar Current (ACC) of an average strength of 130 Sv. The ACC is constrained topographically mainly by Drake passage between Antarctica and South America, which forms a 800-km wide sill. The zonal transport is also partly driven by the sill, which enhances transport through zonal pressure gradients [Gill and Bryan, 1971] and by geostrophic transport due to the meridional temperature gradient [Cai and Baines, 1996]. The momentum input by the wind is understood to be largely dissipated by form stress over rough topography [Munk and Palmen, 1951] transferred downward through a turbulent eddy field [Johnson and Bryden, 1989, Bryden and Cunningham, 2003]. Via Ekman pumping, the westerly winds over the Southern Ocean transport water northwards that forces water from depth up to the surface, causing shoaling of isopycnals (density surfaces). Due to the northward volume transport of water, the Southern Ocean forms a key area in the large scale overturning circulation in the ocean.

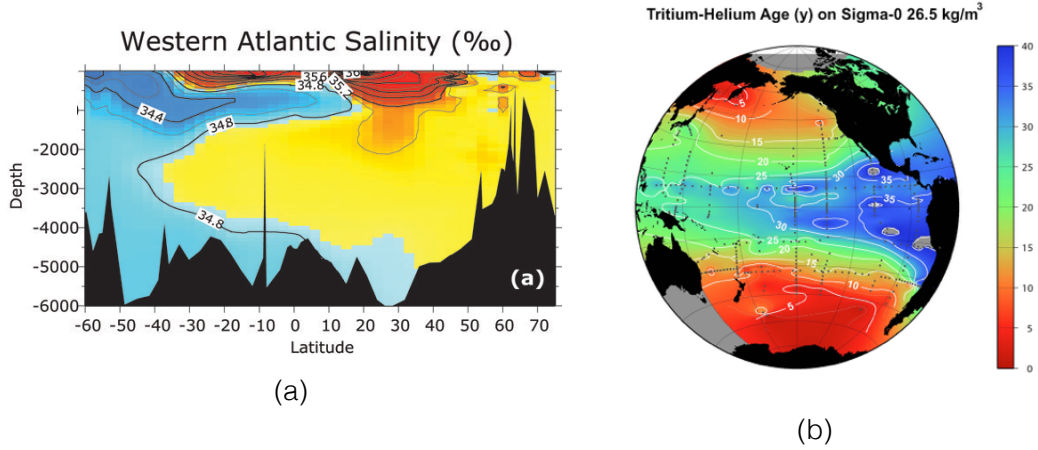


Figure 1.2: a) Atlantic transect of salinity from 1993 [Oppo and Curry, 2012], with lines showing salinity contours and red indicates a high salinity, b) ocean age determined from tritium-helium decay on the 1026.5 kg/m^3 density surface, from the Woods Hole Oceanographic Institute [WHOI, 2015] (the years given are not accurate but are used to represent relative age).

1.1.3 Meridional overturning circulation

Mixing of surface waters with sub-thermocline deep water requires the densification of surface water parcels through temperature and salinity forcing. Similarly, the upwelling of deep water to the surface requires mechanical forcing, so surface-deep water exchange only occurs in specific places, most notably at the high latitudes.

In the Southern Ocean, Ekman pumping upwells deep water along isopycnals (see Fig. 1.3a). Once upwelled, the majority of water is advected northwards and warmed. Where the northward advected water meets warmer and more buoyant water from the subtropical gyre boundary current, it sinks underneath forming Antarctic Intermediate Water, AAIW, which is advected northward underneath the thermocline. AAIW is advected mainly to the North Atlantic where it is mixed into the seasonal thermocline in the sub-polar gyre. South of the ACC, there is strong cooling due to winds and ice formation processes (most notably in the Ross and Weddell seas) such that deep convection and vertical mixing are triggered in winter time [Gordon, 1978, Killworth, 1983, Orsi et al., 2002].

Deep waters created South of the ACC form Antarctic Bottom Water, AABW, and are lightened in the ocean interior through diapycnal mixing and upwelled again in the Southern Ocean (e.g. Watson and Garabato [2006]). Due to both the upwelling and sinking

of waters, there are regions of both ocean heat uptake and release (see Fig. 1.3a). Similarly, carbon from deep waters is brought to the surface through upwelling along the ACC, while North and South of the ACC, carbon is removed from the surface and transformed into mode waters³ (e.g. Fig. 1.6). The result is a complex pattern of surface fluxes that exert strong controls on the atmosphere-ocean balance of heat and carbon [e.g. Toggweiler, 1999, Kuhlbrodt et al., 2007, Sigman et al., 2010, Skinner et al., 2010]. In the Northern high latitudes, deep water formation occurs in the Arctic Ocean and Labrador Sea [Killworth, 1983, Rudels et al., 1991, Dickson and Brown, 1994] (see Fig. 1.3b). Strong surface cooling in winter increases the density and leads to deep mixed layer, vertical mixing, convection and deep water formation in the sub-polar and polar regions.

Due to the Atlantic Ocean’s high salinity, the water formed in the Labrador Sea and Arctic Ocean, North Atlantic Deep Water, NADW, has a characteristically high salinity (see Fig. 1.4a). The combination of deep convection and wind-induced divergence in the sub-polar gyre results in a strong connection between the surface and deep ocean and therefore is key in both ventilating sub-thermocline nutrients, as well as storing heat and carbon in the deep ocean (e.g. Levitus et al. [2000], Sabine et al. [2004], Boccaletti et al. [2005], Williams et al. [2011], Haekkinen et al. [2013]). Figure 1.4a shows observed salinity in the Atlantic Ocean, and accordingly shows the distribution of the different water masses of NADW, AAIW and AABW. Figure 1.4b shows the transport of water masses in the Atlantic basin more simplistically and serves as a representation for the global overturning circulation. We make a distinction between the bottom overturning cell, containing AABW, and the upper overturning cell, formed mainly of NADW and AAIW, often referred to as the Meridional Overturning Circulation, MOC. Despite the distinction, the two cells are by no means separate since water masses are mixed in the interior [Talley, 2013]. The global overturning circulation is a complex system and while surface winds and thermohaline forcing are important drivers, they are certainly not the only ones. Other processes such as diapycnal mixing and diffusion in the ocean interior and Southern Ocean eddies are key components of the global mechanism [Wunsch and Heimbach, 2013]. The ocean dynamics are non-linear, where advective processes often dominate viscous ones.⁴ Particularly high eddy kinetic energies are observed in regions of strong rotation, vertical shear and density gradients such as in ocean fronts [Orlanski and

³Mode waters are sub-surface water masses with characteristic temperature and salinity properties.

⁴Reynolds numbers in the ocean, given the average current speed and basin length scales can be on the order of 10^{11} .

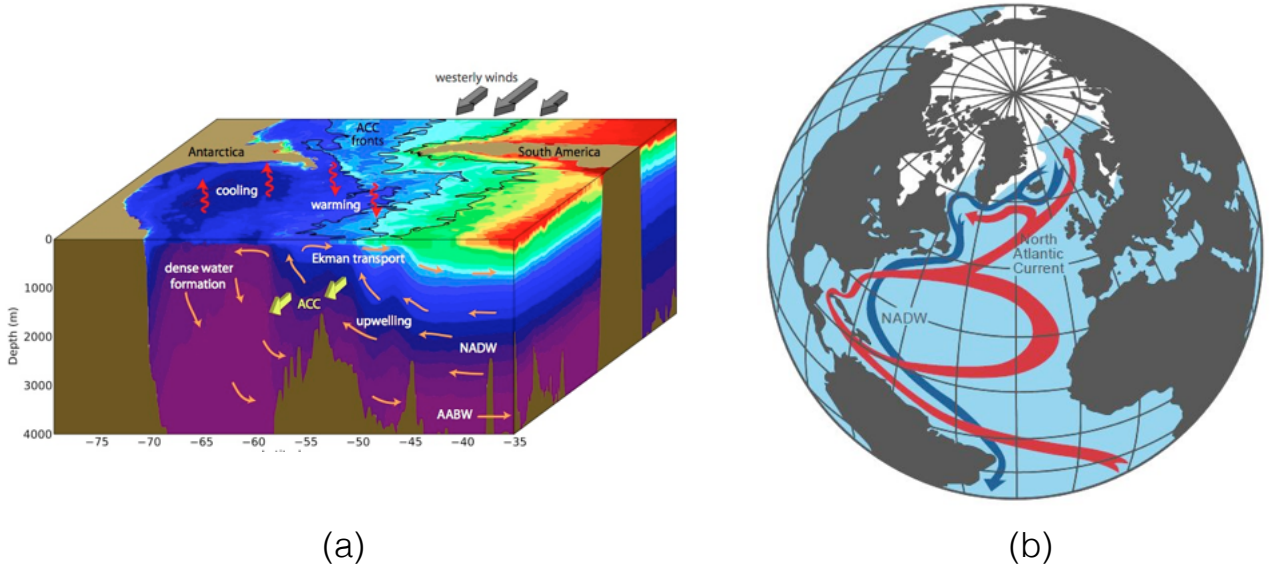
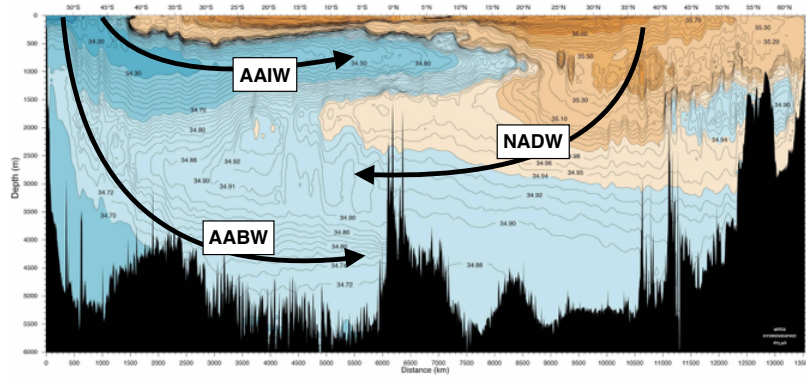


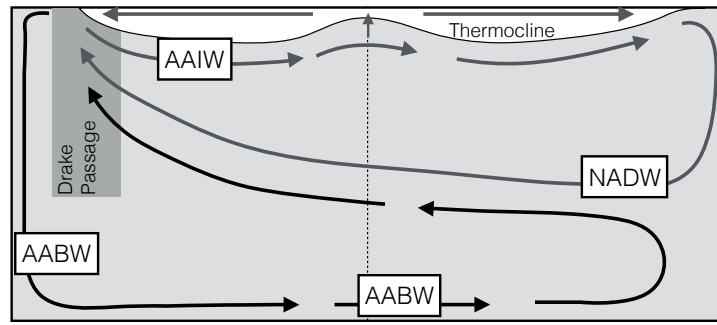
Figure 1.3: a) Diagram of the Southern Ocean circulation, with the colours showing temperature, from the Australian National University [ANU, 2015], b) diagram of the simplified North Atlantic overturning circulation, with red showing warm surface waters and blue showing colder deep waters in the southward return current. The transformation of warm (red) to cold (blue) occurs in the Labrador Sea and the Arctic Ocean, from Rahmstorf [1997].

Cox, 1973]. Such regions include the Gulf Stream, Kuroshio Current, equatorial currents and the ACC. In the context of the large scale circulation, eddies are especially important in the Southern Ocean and their role within the overturning has been described using residual-mean theory [Marshall and Radko, 2003].

The upwelling of dense water and subsequent tilting of isopycnal in the Southern Ocean raises the potential energy of the system. The density gradients induced by the flow create baroclinic instabilities resulting in a mesoscale eddy field that acts to flatten the isopycnal, effectively extracting potential energy from the baroclinic field and transporting mass southward at the surface. The Ekman-induced northward surface velocities imply a mean Eulerian overturning cell, $\bar{\Psi}$, known as the Deacon cell [Doos and Webb, 1994]. However, the eddy field acts to transport mass in the opposite sense, implying an opposing velocity field (the “bolus” velocity field) and opposing cell, Ψ^* . The total transport is given by the sum of the two cells: $\Psi = \bar{\Psi} + \Psi^*$, the residual circulation (see Fig. 1.5). Since mesoscale eddies are 10 to 100 km in diameter, they are largely unresolved in coarse general circulation models, so are accounted for by a bolus velocity parametrization that depends on the sloping of isopycnals [Gent and McWilliams, 1990].



(a)



(b)

Figure 1.4: a) Salinity (in practical salinity units) measured along ship track along 25W in the Atlantic Ocean, where orange indicate high values and cyan low values (from Talley [2014]), b) Schematic of the meridional overturning circulation (adapted from Fig. 2.4.10 in Sarmiento and Gruber [2006])

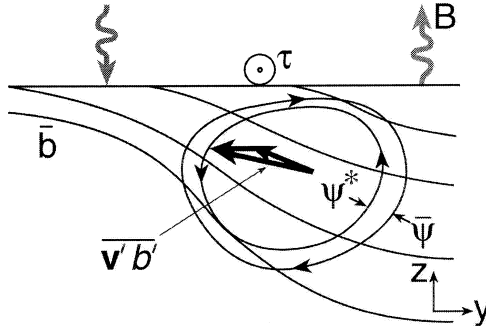


Figure 1.5: Residual mean theory in the Southern Ocean: the residual circulation comprises the sum of the clockwise Eulerian $\bar{\Psi}$ circulation, driven by surface wind stress τ and surface buoyancy forcing $\mathbf{B}$, which steepens the isopycnals $\bar{b}$. The anti-clockwise bolus circulation Ψ^* is driven by eddies and acts to flatten the isopycnals. The eddy field drives a southward buoyancy flux $\overline{v'b'}$. Figure from Marshall and Radko [2003], Lauderdale [2010].

The strength of the eddy field and bolus velocity depends on the tilting of isopycnals, so it follows that as westerly winds over the Southern Ocean increase and the isopycnals

are tilted more, the eddy response strengthens to flatten the isopycnals. The presence of eddies therefore reduces the sensitivity of the Southern Ocean density field to wind stress and buoyancy perturbations. The consequential reduction in sensitivity of the ACC and Ekman transport with respect to wind stress perturbations is known as eddy saturation or eddy compensation [Straub, 1993, Hallberg and Gnanadesikan, 2001]. It is generally found in eddy-resolving models that the ACC’s strength is nearly insensitive to wind stress perturbations, although the global residual overturning circulation is only partially compensated [Dufour et al., 2012, Meredith et al., 2012, Munday et al., 2013].

1.2 Ocean biogeochemistry

The physical processes described have a defining role in setting the distribution of carbon on the world’s ocean. The physical transfer of carbon between the atmosphere and surface ocean and between the surface and deep ocean is largely determined by physical processes, however the biological transfer of carbon is also affected by the advected nutrient field. The details of the ocean carbon distribution are described below.

1.2.1 The air-sea carbon flux

Carbon anomalies enter the ocean via surface exchange with the atmosphere. The transfer of CO_2 from the atmosphere to the ocean occurs through turbulent and diffusive exchange as well as bubbles formed by breaking waves. The process depends strongly on wind speed and many empirical relationships between wind speed and the gas transfer velocity have been proposed [Higbie, 1935, Liss and Merlivat, 1986, Tans et al., 1990, Erickson, 1993, Wanninkhof, 1992, Wanninkhof and McGillis, 1999, Nightingale et al., 2000, McGillis et al., 2001b,a]. The most common parametrization of the air-sea carbon flux, $F_{air-sea}$, follows Wanninkhof [1992] and takes the form:

$$F_{air-sea} = k(p\text{CO}_2^a - p\text{CO}_2^o), \quad (1.1)$$

where $k = 0.39u_w^2 (Sc/660)^{-1/2}$ and $p\text{CO}_2^a$ and $p\text{CO}_2^o$ are the atmospheric and ocean partial pressure of CO_2 respectively, k is the gas transfer velocity, u_w is the surface wind speed and Sc is the Schmidt number which quantifies the thickness of the mass transfer boundary layer (the constant 0.39 is determined empirically and 660 is the Schmidt number for seawater at 20°C). The air-sea flux of carbon is largely determined by the difference in atmospheric and ocean partial pressure. Since atmospheric CO_2 is generally well-mixed,

the broad-scale pattern of the air-sea flux is determined by the ocean $p\text{CO}_2$ and the air-sea heat flux.

Figure 1.6a-b shows the result of gridded in-situ measurements of ocean $p\text{CO}_2$ in 1990 in the form of the air-sea $p\text{CO}_2$ difference in Boreal summer (panel a) and winter (panel b). The pattern of sea-air heat flux is shown in panel c, and the yearly mean of sea-air CO_2 flux in 1990 is shown in Fig. 1.6d. The pattern of the sea-air CO_2 flux is seen to be largely dominated by the air-sea $p\text{CO}_2$ difference with the sea-air heat flux playing an important role at high latitudes.

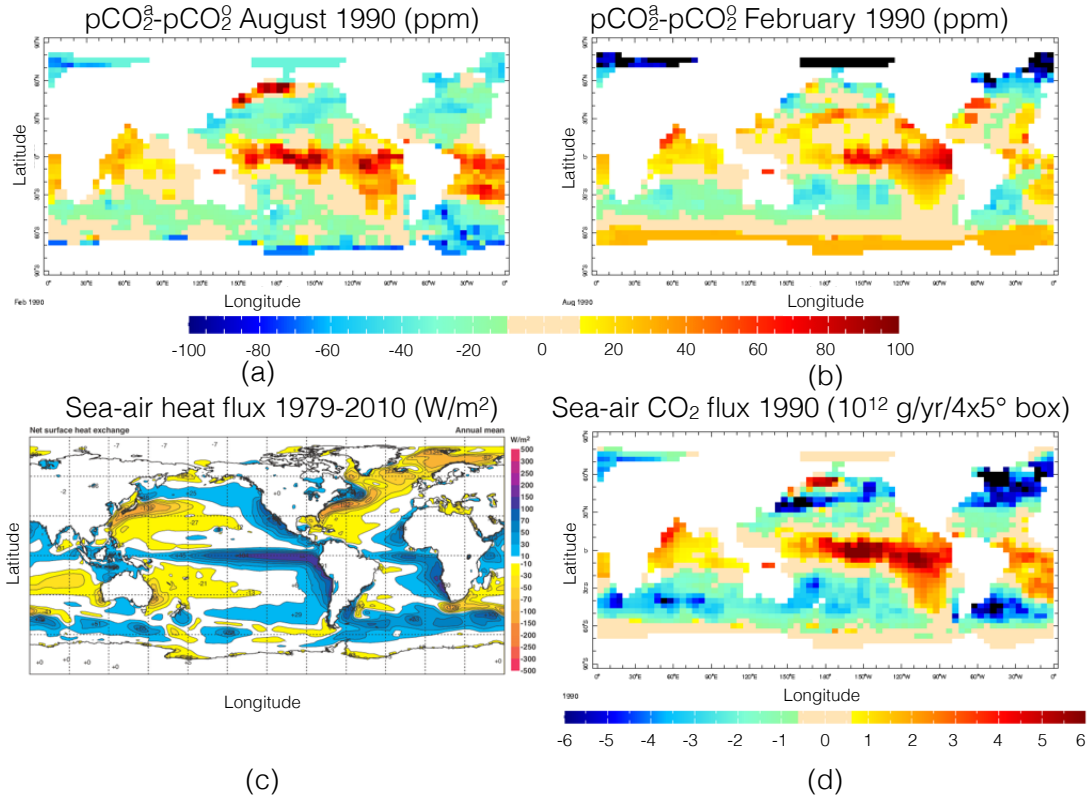


Figure 1.6: Observed yearly mean atmosphere-ocean $p\text{CO}_2$ difference in 1990 (ppm) in a) August and b) February from data [Takahashi, 2015]; c) mean sea-air heat flux between 1979 and 2010 (W/m^2), from ERA-40 re-analysis [Uppala et al., 2005]; d) Yearly mean sea-air CO_2 flux in 1990 ($10^{12} \text{ g/yr/4x5}^\circ \text{ box}$) from data [Takahashi, 2015].

Ocean $p\text{CO}_2$ is generally high in the Tropics since gases are less soluble at high temperatures.⁵ Accordingly, ocean $p\text{CO}_2$ is higher than atmospheric $p\text{CO}_2$ causing ocean CO_2 outgassing. Water parcels cool as they are advected to the poles along western boundary currents, so $p\text{CO}_2$ decreases and drives a flux of carbon from the atmosphere into the

⁵The ability of water to dissolve CO_2 molecules, the solubility, decreases with temperature and increases with salinity. Ocean carbon solubility is generally driven by temperature.

ocean. The pattern of ocean CO_2 outgassing in the Tropics and uptake at high latitudes determines to a large extent the distribution of carbon in the ocean interior, alongside sub-surface ocean dynamics.

1.2.2 The components of the ocean carbon cycle

Once dissolved in sea water, dissolved CO_2 , CO_2^* , interacts with water molecules to form bicarbonate HCO_3^- and carbonate ions CO_3^{2-} . The total concentration of carbon, or Dissolved Inorganic Carbon (DIC), is given by:

$$\text{DIC} = [\text{CO}_2^*] + [\text{HCO}_3^-] + [\text{CO}_3^{2-}], \quad (1.2)$$

where square brackets indicate concentration. While numerical models solve for the individual chemical components (see section 2.3), to better understand the ocean carbon cycle and its response to forcings, DIC is often broken down into two main components, the preformed, C_{pre} , and regenerated, C_{reg} , carbon concentrations [Broecker, 1974, Broecker and Peng., 1982, Broecker and Takahashi, 1985, Toggweiler et al., 2003b]. The preformed component is the carbon concentration of a parcel of water when last in contact with the atmosphere. The regenerated component is the accumulated DIC due to biological processes after subduction.

The preformed carbon can be decomposed further into the concentration the water parcel would have if it was at equilibrium with the atmosphere, the saturated carbon concentration C_{sat} and the disequilibrium concentration between the ocean and atmosphere at the time of subduction, C_{dis} (i.e. the difference between C_{pre} and C_{sat}). C_{sat} is a function of temperature, salinity, alkalinity (it is assumed these are conserved in the ocean interior) and the atmospheric $p\text{CO}_2$. C_{sat} also comprises contributions from anthropogenic atmospheric CO_2 anomalies, however, C_{sat} is generally defined with respect to pre-industrial atmospheric CO_2 levels. The anthropogenic contribution is therefore captured in a separate component, C_{ant} .

The regenerated carbon concentration can be separated into contributions from the respiration of soft organic tissue, C_{soft} and from the remineralization of calcium carbonate shells, C_{carb} . Respiration of organic matter utilises oxygen such that C_{soft} can be calculated from the Apparent Oxygen Utilization (AOU), the measured undersaturation

of oxygen⁶: $C_{soft} = -r_{C:O}AOU$, where $r_{C:O}$ is the Redfield ratio between carbon and oxygen⁷ [Redfield, 1934]. DIC can therefore be written as:

$$DIC = C_{sat} + C_{dis} + C_{soft} + C_{carb} + C_{ant}. \quad (1.3)$$

To illustrate the contributions of each of the terms, Fig. 1.7 shows the zonal average in the Atlantic of total DIC (Fig. 1.7a) and each of the components C_{sat} , C_{dis} , C_{soft} , C_{carb} from a realistic ocean-only pre-industrial control simulation such that $C_{ant} = 0$ (the calculation of the components requires preformed and regenerated alkalinity and oxygen data fields, which are easily available in models but not in observations). The results are the following:

- Atmospheric CO₂ is generally well-mixed such that the distribution of C_{sat} concentration (Fig. 1.7b) broadly coincides with the density structure (Carbon is more soluble at low temperatures and high salinities [Volk and Hoffert, 1985]).
- C_{dis} (Fig. 1.7c) represents the over- or under-saturation of water parcels with respect to saturated carbon at the time of subduction. Mode water originating in the Southern Ocean generally has positive C_{dis} concentrations, since the equilibration timescale of ocean carbon with the atmosphere is shorter than the residence time of water parcels [Ito and Follows, 2013]. Deep biological carbon-rich water upwelled in the Southern Ocean has insufficient time to outgas the excess carbon and equilibrate with the atmosphere before subduction, so it has a positive C_{dis} concentration. The re-subducted regenerated carbon is therefore re-labeled as C_{dis} . Conversely, in the North Atlantic, water parcels are in equilibrium with the atmosphere prior to being advected to the high latitudes. In the sub-polar region, biological activity strips carbon from the surface waters before deep water formation and re-equilibration with the atmosphere, resulting in negative C_{dis} concentrations.
- C_{soft} accumulates with time, leading to higher concentrations in older waters. C_{soft} is therefore high in the Antarctic Bottom Water as shown in Fig. 1.7d. Due to equatorial upwelling over the Eastern tropical shadow zone, C_{soft} also accumulates below the mixed layer in equatorial regions. Upon remineralisation of soft tissue, waters are stripped of dissolved oxygen, causing low oxygen concentrations leading to oxygen minimum zones (OMZ) [Cline and Richards, 1972, Paulmier and Ruiz-Pino,

⁶Oxygen has an air-sea equilibration timescale of less than a month, so the oxygen concentration of subducted water parcels are assumed to be equilibrium with the atmosphere.

⁷The Redfield ratio is the stoichiometric ratio between carbon, nitrogen, phosphate and oxygen in organic matter

2009, Stramma et al., 2012]. Measurements of the extent of OMZs in the Pacific and Atlantic have shown these zones increasing over the past decades [Keeling et al., 2010, Stramma et al., 2012] and models also project further expansions of the OMZs due to anthropogenic climate change [Sarmiento et al., 1998, Oschlies et al., 2008].

- The depth of carbonate remineralisation is deeper than for soft organic tissue [Tsunogai and Noriki, 1991, Anderson and Sarmiento, 1994]. The majority of carbonate is therefore remineralised deep in the water column, showing a maximum in the AABW cell (Fig. 1.7e).

Overall, 93.5% of DIC is stored as saturated carbon, 4.5% as C_{soft} , 1.5% as C_{carb} and 0.5% as C_{dis} [Williams and Follows, 2011].

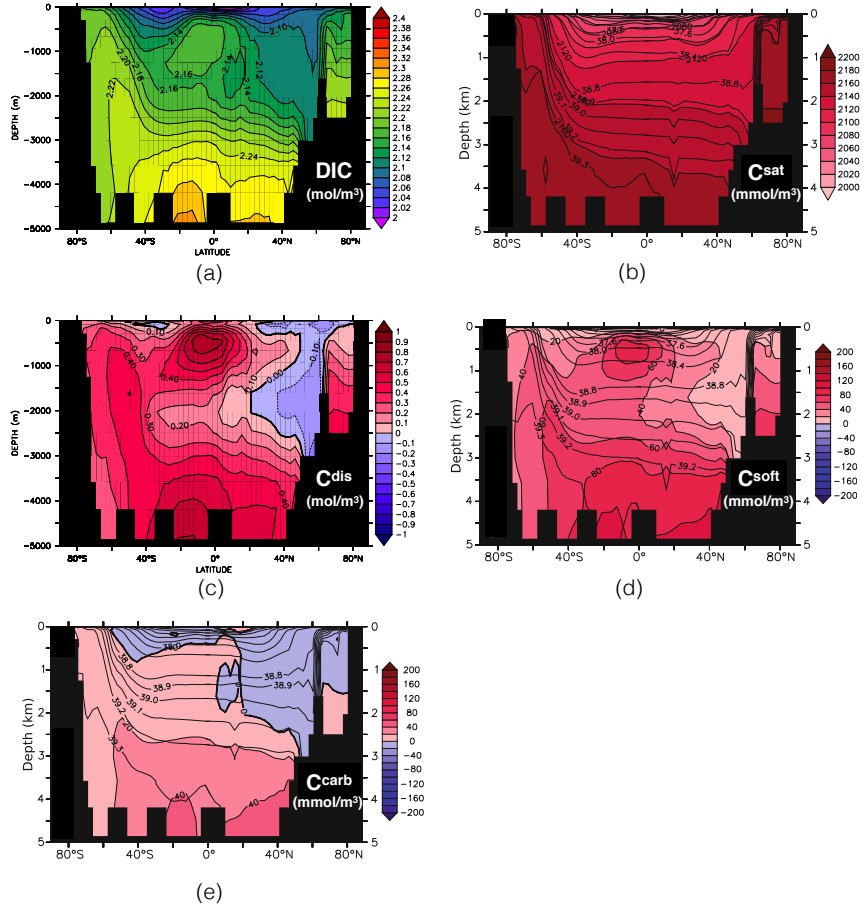


Figure 1.7: Zonal averages of a) dissolved inorganic carbon, DIC, (colours) and potential density (contours, in kg m^{-2}), b) saturated carbon C_{sat} , c) disequilibrium carbon C_{dis} , d) soft tissue carbon C_{soft} and e) carbonate C_{carb} concentrations in the Atlantic basin from a realistic bathymetry ocean-only pre-industrial control simulation. a)-b) are taken from Lauderdale [2010] and c)-e) from Lauderdale et al. [2013] (note the different units).

The modelled DIC distribution shown in Fig. 1.7a matches recent observations of zonal mean Atlantic DIC North of 40°S (Fig. 1.8a). The total DIC comprises both natural and anthropogenic components. However the anthropogenic carbon concentrations C_{ant} are estimated to be fairly small, generally less than 60 mmol/m³ (Fig. 1.8b). Since anthropogenic carbon has only been measured in the atmosphere since the industrial revolution in 1850, C_{ant} is generally observed in the upper 1000m of the ocean. Methods for estimating C_{ant} vary: in the past, the ΔC^* tracer method [Gruber et al., 1996] was common⁸ but recently other methods such as Green's function analysis have been used to map surface signals into the ocean interior [Khatiwala et al., 2009].

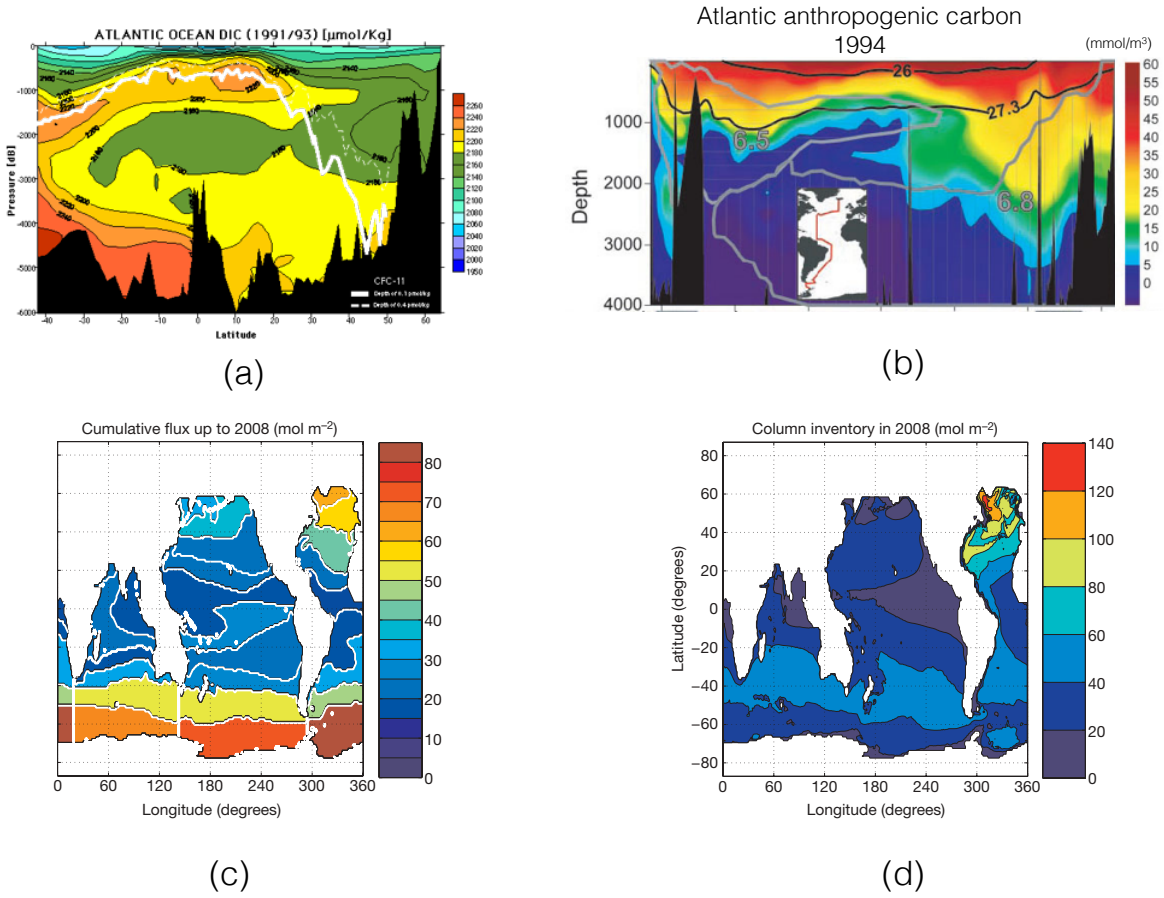


Figure 1.8: a) Zonal mean Atlantic total DIC in 1991-1993 from observations [NOAA, 1993], b) Zonal mean Atlantic anthropogenic carbon concentration in 2004, estimated using the ΔC^* method [Sabine et al., 2004], with black lines showing density contours, c) Cumulative air-sea flux of C_{ant} in 2008 since 1850 [Khatiwala et al., 2009], d) total column inventory of C_{ant} in 2008 since 1850 [Khatiwala et al., 2009] using Green's functions.

⁸ ΔC^* is calculated as the residual between the measured DIC in the ocean and the contributions from all other reservoirs: $\Delta C^* = \text{DIC} - C_{sat} - C_{soft} - C_{carb} - C_{dis}$. In this case, C_{dis} is estimated using passive tracer measurements along isopycnal surfaces [Sabine and Tanhua, 2010].

Anthropogenic carbon is removed from the mixed layer due to the mixing between surface and deep waters in the North Atlantic sub-polar gyre and Southern Ocean. As a result, the removal of mixed layer C_{ant} in these areas leads to a strong cumulative air-sea flux of anthropogenic carbon (Fig. 1.8c). In the Southern Ocean, due to the lateral advection of Antarctic Intermediate Water, the anthropogenic carbon is not stored locally in the water column, unlike in the North Atlantic where the deep convection and mixing cause C_{ant} to accumulate in the water column (Fig. 1.8d) [Khatiwala et al., 2009].

1.3 Ocean-carbon feedbacks

1.3.1 Carbon pumps

An important controlling factor in the atmosphere-ocean balance of carbon is the global vertical gradient of DIC, since DIC concentration increases with depth. The more carbon is stored at depth, the lower atmospheric CO_2 . The physical and biological processes setting the vertical DIC gradient in the ocean are known as the solubility and biological carbon pumps, respectively [Volk and Hoffert, 1985].

The solubility pump:

The solubility pump refers to the physical storage of carbon at depth and is controlled by the vertical temperature gradient. Patterns of ocean temperature therefore play an important role in the solubility pump through the C_{sat} reservoir. Increases in ocean temperatures reduce ocean carbon solubility which reduces the physical storage of carbon at depth and therefore increases atmospheric CO_2 [Omta et al., 2011, Goodwin and Lenton, 2009]. The solubility pump can therefore act as a positive feedback mechanism on atmospheric CO_2 : atmospheric CO_2 emissions strengthen the greenhouse effect, thereby raising atmospheric and surface ocean temperatures, which decrease ocean carbon solubility and therefore increase atmospheric CO_2 further. The largest uncertainty in the strength of solubility pump feedback is the relative importance of low and high latitudes, mainly due to the inadequacies of models in representing local air-sea fluxes, along with ocean convection and mixing [Broecker et al., 1999, Toggweiler et al., 2003a, Marinov and Gnanadesikan, 2011].

Physical mechanisms also contribute towards the solubility pump, particularly large scale ocean circulation [Schwinger et al., 2014, Randerson et al., 2015]. Studies such as

Goodwin and Lenton [2009] suggest that dynamically-induced changes in ocean circulation in response to anthropogenic forcing in transient runs (such as a reduction in the Atlantic MOC) have an impact of similar magnitude on atmospheric CO_2 as does direct warming. Atmospheric CO_2 changes induced by changes in the overturning circulation are generally a result of changes in the C_{dis} ocean carbon reservoir [DeVries and Primeau, 2009]. Furthermore, an increase in wind-driven surface currents can amplify the importance of low latitudes by advecting mid-latitude properties throughout the thermocline [Follows et al., 1996, 2002].

The biological pump:

Biological activity in the ocean surface extracts carbon from the mixed layer and stores it in the deep ocean, thereby lowering mixed layer ocean $p\text{CO}_2$ and increasing the air-sea CO_2 flux. As carbon is removed from the surface ocean through biological activity, surface ocean $p\text{CO}_2$ is reduced which drives a flux of carbon from the atmosphere into the ocean, and in doing so lowers atmospheric CO_2 levels. The amount of ocean nutrients cycled through biological activity varies linearly with biological carbon, so if more nutrients are stored in the regenerated versus the preformed pool, atmospheric CO_2 will decrease [Ito and Follows, 2005, Marinov et al., 2008a, Kwon et al., 2011, Honjo et al., 2014, Lopes et al., 2015].

Ito and Follows [2005] define the efficiency of the global biological pump N^* , as the ratio of the global mean regenerated concentration of nutrients, N_b , over the total global mean nutrient concentration (preformed plus regenerated), N : $N^* = \frac{N_b}{N}$. A higher efficiency means lower atmospheric CO_2 concentrations. Adaptations in the biological pump can be incurred through variations in nutrient utilization: more efficient surface biology (through for example, a higher supply of micro-nutrients such as iron) uses more nutrients and increases the regenerated pool of carbon. Previous studies suggest that the biology in the high latitude ocean is dominant in determining global nutrient utilization and atmospheric CO_2 [Sarmiento and Toggweiler, 1984, Siegenthaler and Wenk, 1984, Knox and McElroy, 1984, Sigman and Boyle, 2000].

In response to climate change, dynamical adjustments in the ocean play an important role in the biological pump by re-distributing regenerated carbon and nutrients. In the Southern Ocean, there is a separation between the formation of Antarctic deep water (in

the Antarctic region) and intermediate water (in the sub-Antarctic region). The former has a strong direct impact on atmospheric CO₂, whereas, the latter has a stronger control over global biological carbon production, via regulation of upwelling nutrients [Marinov et al., 2006]. Studies such as Lovenduski et al. [2007], Le Quere et al. [2007] and Le Quere et al. [2010] link increases in westerly winds over the Southern Ocean to an increase in upwelling of deep organic carbon, reducing the efficiency of the biological pump and elevating atmospheric CO₂ levels. Such a process can therefore act as a positive feedback on atmospheric CO₂ levels: atmospheric CO₂ emissions strengthen atmospheric temperature gradients and surface winds, which upwell deep organic carbon that is outgassed into the atmosphere, thereby increasing atmospheric CO₂ further. However, the current and future magnitudes of this effect are uncertain.

1.3.2 Quantifying climate feedbacks

The global net carbon cycle feedback on atmospheric CO₂ levels is believed to be positive [Fung et al., 2005, Friedlingstein et al., 2001, Schwinger et al., 2014, Randerson et al., 2015]. On a global scale, Le Quere et al. [2010] estimate that climate change-induced warming and shifts in wind patterns have reduced the current ocean carbon uptake by 12%. Plattner et al. [2001] suggest that a decrease in ocean CO₂ uptake of 5% to 16% will last until the end of the 21st century. Using an ensemble of carbon climate models, Friedlingstein et al. [2006] estimate that the net climate change-induced feedback (due to both increased atmospheric CO₂ and temperature) could further increase global temperatures by 0.1 to 1.5°C by 2100.

The feedback onto atmospheric $p\text{CO}_2$ can be separated into two processes: a concentration-climate feedback associated with the biogeochemical response to increased atmospheric CO₂ concentrations and a climate-carbon feedback due to secondary climate change effects such as increased global mean temperatures [Gregory et al., 2009]. Gregory et al. [2009] demonstrate that the strength of the transient feedbacks are scenario-dependent, but that the concentration-carbon feedback is generally 4 times larger than the climate-carbon feedback on atmospheric $p\text{CO}_2$ (owing mostly to the increase in ocean uptake of atmospheric CO₂ as atmospheric CO₂ levels increase, due to the form of the air-sea carbon flux, eqn. 1.1). Model studies such as Schwinger et al. [2014] show that the climate-carbon feedback could reduce ocean carbon uptake by 100 Pg of C (or roughly 50 ppm) after increasing atmospheric $p\text{CO}_2$ levels by 1% a year for 140 years. Comparison studies between

climate- and concentration-feedback studies are generally done by only coupling either the radiative or biogeochemical forcing of carbon emissions. The more mechanistic studies of the ocean carbon feedback focus on the climate-carbon feedback and aim to calculate the change in atmospheric carbon due to a given change in ocean carbon inventory [Goodwin et al., 2007, Schmittner et al., 2007] or air-sea carbon flux [Lovenduski et al., 2007, Le Quere et al., 2007].

Goodwin et al. [2007] calculate the atmospheric feedback of a change in total ocean carbon. The authors find that the atmospheric partial pressure of CO_2 , $p\text{CO}_2^a$, scales exponentially with a change in total ocean carbon ΔC , such that

$$p\text{CO}_2^a = p\text{CO}_{2i}^a e^{\Delta C/I_B}, \quad (1.4)$$

where $p\text{CO}_{2i}^a$ is the initial atmospheric $p\text{CO}_2$ and I_B is the buffered carbon inventory (defined below). Using eqn. 1.4, Goodwin et al. [2008] show that ΔC can be split into different carbon cycle components' contributions such that

$$\Delta C = V (\Delta C_{sat} + \Delta C_{dis} + \Delta C_{soft} + \Delta C_{ant}),^9$$

where V is the volume of the ocean. A change in atmospheric $p\text{CO}_2$ due to a change in a given ocean variable can be estimated by applying the chain rule. For example, to calculate the atmospheric feedback due ocean temperature change ΔT via ΔC_{sat} , we can write $\Delta C_{sat} = \frac{\partial C_{sat}}{\partial T} \Delta T$. While the sensitivity of the saturated carbon reservoir to temperature has been well-defined, the sensitivity of the components C_{reg} , C_{dis} and C_{ant} to warming (e.g. $\frac{\partial C_{reg}}{\partial T}$), and particularly warming patterns affecting ocean dynamics and biogeochemistry, are uncertain.

An important concept in the calculation of the atmospheric feedbacks is the buffered carbon inventory, I_B . The buffered carbon inventory is the total amount of carbon in the air-sea system that is available for exchange:

$$I_B = I_A + \frac{I_O}{R}, \quad (1.5)$$

where I_A and I_O are the total amount of carbon in the atmosphere and ocean, respectively. R is the Revelle Buffer Factor [Revelle and Suess, 1957] defined as

$$R = \frac{\delta p\text{CO}_2}{p\text{CO}_2} / \frac{\delta \text{DIC}}{\text{DIC}}. \quad (1.6)$$

⁹Inclusion of a ΔC_{carb} term is slightly more complicated due to effect of C_{carb} on alkalinity, see Goodwin et al. [2007].

R measures the fractional ocean $p\text{CO}_2$ change given a fractional change in DIC and vice versa. R quantifies the amount of carbon that can be stored as DIC for a given change in atmospheric CO_2 concentrations. A high buffer factor system means DIC would not change significantly for a given change in $p\text{CO}_2$ and as a result, more of the excess carbon remains in the atmosphere. Inversely, for a small buffer factor, the ocean would be able to absorb larger amounts of excess atmospheric carbon. The mean pre-industrial ocean surface R is around 10. Since R depends on the chemical properties of the carbon species, it is highly spatially variable. In the ocean, the value of R varies from 8 in the Tropics to 16 in high latitudes, which reveals that the Tropics are more capable of absorbing atmospheric carbon anomalies due to local chemistry (see the higher concentrations of C_{ant} at lower latitudes in Fig. 1.8b). The quantity R is therefore useful to estimate how much carbon a particular region of the ocean can absorb and to quantify local climate-carbon feedbacks.

Goodwin et al. [2009] and Goodwin and Ridgwell [2010] follow the framework developed in Goodwin et al. [2007] which allows the climate impact of global scale changes in ocean chemistry to be calculated without the explicit calculation of climate feedback parameters. Goodwin et al. [2009] calculate the radiative forcing change due to the impact of anthropogenic carbon emissions on the chemical partitioning of carbon in the ocean and the buffered carbon inventory. The current ocean chemistry means the climate is more sensitive to carbon emissions now than in the past 400 millions years, such that a 1000 Pg of C carbon emission increases global radiative forcing by $1.5\text{--}1.7 \text{ Wm}^{-2}$ as opposed to less than 0.5 Wm^{-2} 200 million years ago. Goodwin and Lenton [2009] calculate the atmospheric CO_2 change due to a global increase in ocean heat uptake, reducing ocean carbon solubility. They find that the ocean outgassing due to solubility increased from $0.08 \pm 0.01 \text{ PgC yr}^{-1}$ to $0.19 \pm 0.02 \text{ PgC yr}^{-1}$ from 1961 to 2003, however the approach does not account for ocean carbon outgassing due to dynamical changes induced by surface warming.

More process-based studies focus on atmospheric CO_2 changes due to adjustments in ocean dynamics. In particular, high latitude regions have been the focus of recent studies. Increasing westerly winds over the Southern Ocean might affect the atmosphere-ocean distribution of heat and carbon [Le Quere et al., 2007, Lovenduski et al., 2008, Lovenduski and Ito, 2009], via Ekman pumping and a decrease in the stratification in the Southern Ocean. The Southern Ocean is a place where the deep carbon-rich ocean is ventilated, so

an increase in Westerly winds is generally agreed to imply a transfer of carbon from the ocean into the atmosphere [Sarmiento et al., 1998, Matear and Hirst, 1999, Caldeira and Duffy, 2000] which will further elevate atmospheric CO₂ levels.

In the contemporary climate, the Southern Ocean acts as a net carbon sink due to the high levels of atmospheric CO₂. Le Quere et al. [2007] use observations of atmospheric CO₂ and atmospheric inversions to find a negative trend in the Southern Ocean sink of atmospheric carbon between 1981 and 2004 with an overall reduction in uptake of carbon of 0.08 Ptg of carbon per year over this period. They attribute this reduction to the observed increase in westerly winds. Additionally, latitudinal shifts in the maximum wind stress will affect the upwelling of carbon. A poleward shift is likely to reduce the effect of decreasing stratification, though such effects are found to be secondary to those of the wind strength change [Russell et al., 2006, Lauderdale et al., 2013].

In the North Atlantic, studies such as Ullman et al. [2009] and McKinley et al. [2011] find an increase in the sub-polar gyre carbon sink driven by vertical mixing between 1992 and 2006, using both models and observations. McKinley et al. [2011] note that the trend in the sink depends on the time-frame of the study, since they discern no trend when lengthening the period under study from 15 years to 25. They conclude that the sub-polar gyre is affected by oscillatory modes of atmospheric forcing on decadal timescales, such as the North Atlantic Oscillation which drives zonal winds across the region. Observational uncertainties limit the conclusions that can be drawn however, especially given the short time series that are available.

Other studies such as Perez et al. [2013] however, find a decrease in the North Atlantic sub-polar gyre carbon sink between 1990 and 2006, which they attribute to a weakening in the meridional overturning circulation (MOC) leading to weaker sea-air heat fluxes out of northward advected water parcels. Further South, Ullman et al. [2009] and McKinley et al. [2011] note that the anthropogenic-induced warming causes a weakening in the sub-tropical gyre carbon sink due to a reduction of the solubility of carbon. The evolution of the sub-polar and sub-tropical gyres has therefore only been studied on decadal timescales and while it is clear that dynamical processes such as wind-induced mixing and the MOC play a role in the carbon sink trend, the net effects are uncertain even on these short

timescales. The uncertainties in the North Atlantic carbon sink on the longer centennial timescales associated with climate change are therefore even larger.

1.4 Thesis aims

While climate-carbon feedbacks of the ocean carbon cycle on atmospheric CO₂ have been well studied, many gaps in our understanding still remain due to the multitude of processes and global interactions involved. As a whole, this thesis aims to disentangle ocean carbon feedbacks in terms of region and mechanism, as well as provide an outlook to how such feedbacks may evolve as the climate changes. In Chapter 2, we firstly use a simple 1-D model to motivate the need for a thorough understanding of how air-sea heat and carbon fluxes evolve in the changing climate after which the tools used to tackle our aims are introduced.

Studies such as Le Quere et al. [2007], Lovenduski et al. [2008] and Lauderdale et al. [2013] aim to quantify the feedback of surface wind stress, in the Southern Ocean in this case, through modelling studies without an outlook as to how feedback may change as the ocean warms or the biogeochemistry adapts. In chapter 3, we quantify the atmospheric CO₂ feedback of Southern Ocean winds and project how the feedback may change in the future.

Changes in Southern Ocean winds also affect global ocean dynamics and the carbon cycle outside of the Southern Ocean, which will provide a secondary feedback. The secondary ocean-carbon feedback due to Southern Ocean wind stress will be addressed in chapter 4.

In chapter 5, we build on studies such as Goodwin et al. [2007, 2009], Goodwin and Lenton [2009], Goodwin and Ridgwell [2010] by decomposing the response of the ocean carbon cycle to warming, both spatially and mechanistically. Due to their different role in the global ocean circulation, Northern and Southern high latitude carbon will respond differently to warming, as will the mid-latitudes. Each region will therefore play a different role in the natural and anthropogenic exchange of carbon between the ocean and atmosphere. The ocean response to warming is decomposed beyond the solubility-driven

response as studied in Goodwin and Lenton [2009] such that we can understand the role and magnitude of dynamical and biogeochemical feedbacks due to warming.

While ocean carbon storage is closely related to heat storage, we wish to quantify the similarities of ocean carbon and heat uptake in the changing climate. Simple 1-D models often assume similar forms of ocean uptake for both so we use the experiments from the carbon feedback studies to quantify the similarities between ocean heat and carbon storage and where differences arise. The heat uptake in relation to carbon uptake will therefore be discussed in chapter 5 before concluding and discussing how the research presented in the thesis can be applied to future studies.

The overall aims of the thesis can therefore be summarised as follows:

- To provide a theoretical understanding of the sensitivity of atmospheric CO₂ to high latitude wind stress and global (both high- and mid-latitude) ocean warming;
- To disentangle the local and non-local regional contributions of temperature and wind stress feedbacks on natural and anthropogenic carbon;
- To quantify the similarities of ocean heat and carbon uptake in a changing climate.

Motivation and Methods

This chapter illustrates the need to investigate ocean carbon and heat uptake and introduces the tools used. Firstly, a 1-D model is used to motivate the need for an understanding of how the mechanisms and timescales of ocean carbon and heat uptake adjust to climate change (section 2.1). The tools used to conduct the study are the global atmosphere-ocean coupled model HadGEM2-ES (section 2.2) and an idealised MITgcm experimental set-up used to perform perturbation experiments (section 2.3). We introduce a mixed layer carbon budget equation and atmospheric CO₂ feedback equation in section 2.4. These equations will be used in subsequent chapters as a tool to understand and predict the response of ocean and atmospheric pCO₂ changes due to perturbations in surface buoyancy and wind stress.

2.1 Air-sea fluxes and global temperatures

The ocean constitutes the largest active reservoir of carbon and heat in the climate system, so the timescales on which it absorbs atmospheric carbon anomalies are crucial in determining future atmospheric temperatures. Simple climate models often use simple parametrizations to represent ocean heat and carbon uptake which the timescales of ocean heat and carbon uptake. By varying these parametrizations, we can manipulate the timescales of uptake and illustrate the impact these timescales have. We use a simple 1-D model to explore the role of ocean timescales in present and future climates and show how these timescales may change in the future.

2.1.1 1-D Model description

The simple model, initially developed by Allen et al. [2009], is retuned using recent observations (see appendix, section 7.1). The model is constructed as a stacked layer model, where the individual carbon components add up to determine the total carbon in the atmosphere. There is one box of carbon, which is subdivided into 3 different components: C_1 , C_2 and C_3 . The component C_3 is the long term atmospheric CO₂ reservoir, which is only increased by emissions above a pre-industrial value of 280 ppm:

$$\frac{dC_3}{dt} = b_3 E \quad (2.1)$$

The constant b_3 is the Revelle Buffer Factor. The other two components of atmospheric CO_2 , C_2 and C_3 represent the interaction of the atmosphere with the land and ocean. C_2 is increased by a fraction of the total emissions and is decreased by advection into the deep ocean:

$$\frac{dC_2}{dt} = b_1 E - b_0 C_2 \quad (2.2)$$

The second term represents a linearized advection process, with a characteristic timescale b_0 of 200 years. The final carbon component C_1 quantifies the atmospheric CO_2 anomalies in exchange with the surface land and surface ocean:

$$\frac{dC_1}{dt} = b_4 E - b_2 \int_0^t \frac{dC_1(t')}{dt'} \frac{dt'}{\sqrt{t-t'}} \quad (2.3)$$

The transfer of CO_2 between the atmosphere and surface is governed by a diffusive term with timescale b_2 . The sum of b_1 , b_3 and b_4 is the level which atmospheric CO_2 would fall to a year after a pulse injection. While $C_1 + C_2 + C_3$ is the total atmospheric concentration, each component also represents an equilibrium acting on a different timescale. C_1 can be thought of as the excess CO_2 over the short-term equilibrium $C_2 + C_3$. The carbon cycle is driven by CO_2 emissions which consist of a prescribed anthropogenic emission profile, and a linear temperature feedback term:

$$E = E_a + b_5 T' \quad (2.4)$$

where T' is the difference between the temperature anomaly T above the pre-industrial value and an exponentially weighted average of T over the preceding 100 years, to reflect the timescale of response of the system. The T' term captures the changes in terrestrial and oceanic carbon storage in response to rising temperatures [Bender, 2003]. The temperature anomaly T is governed by an energy balance equation with a radiative forcing term, due to carbon emissions, a linear feedback and an additional diffusive flux into the surface:

$$a_1 \frac{dT}{dt} = a_3 \ln\left(\frac{C}{C_0}\right) - a_0 T - a_2 \int_0^t \frac{dT(t')}{dt'} \frac{dt'}{\sqrt{t-t'}} \quad (2.5)$$

where a_1 is the heat capacity of a system with a 75m deep ocean mixed layer with 70% global coverage. a_0 is the feedback parameter that controls the equilibrium climate sensitivity. The feedback term has to be negative for the system to be stable. a_2 controls downward diffusion of heat anomalies into the surface layer. The model is forced by yearly CO_2 emissions E , which consists of a historical exponential rise to the year 2000 followed by a mitigation scenario. To investigate the influence of air-sea carbon and heat uptake on atmospheric CO_2 and temperature, we modify the parametrization of air-sea fluxes to

alter the mechanism and timescale over which the exchange takes place. Each diffusive-type expression can be replaced with a linear term proportional to the temperature or carbon anomaly, e.g.:

$$a_2 \int_0^t \frac{dT(t')}{dt'} \frac{dt'}{\sqrt{t-t'}} \rightarrow a_2 T \quad (2.6)$$

$$b_2 \int_0^{t'} \frac{dC_1(t')}{dt} \frac{dt'}{\sqrt{t-t'}} \rightarrow b_2 C_1 \quad (2.7)$$

The air-surface interaction is therefore reduced to a single timescale process, or an ‘advective’ parametrization. With two different parametrizations of air-sea interaction, advective and diffusive, for each of the temperature and carbon anomaly, four different versions of the model are constructed.¹ The Representative Concentration Pathway 2.6 (RCP2.6) CO₂ emission profile which ensures a maximum in radiative forcing of 3 Wm⁻² is used to compare results from the different simple model versions with a number of models from the Coupled Model Intercomparison Project Phase 5 (CMIP5) archive [Taylor et al., 2012]. The original emission profile from Allen et al. [2009] is also used which is comprised of historical emissions between 1850 and 1980, a 1.7% increase per year from 1980 until 2010 after which the percentage is decreased linearly until 2030 to reach a constant rate of -5%/yr decrease in emission, giving a 1 trillion ton total emission of carbon in the year 2500 (Fig. 2.1a and Fig. S3a in Allen et al. [2009]).

Figure 2.1b shows a spread in peak temperature anomaly of 0.3°C (23% of the maximum value) as a result of the changes in the model’s air-sea flux parametrization. The original fully diffusive model from Allen et al. [2009] lies at the upper end while the advective heat flux model versions project lower peak temperatures. The advective models have a faster response and react more rapidly to a turnover in carbon emissions. The spread in peak CO₂ concentration obtained from the different parametrization, shown in Fig. 2.1c, is much smaller than the spread in temperature anomaly and is approximately 5% of the maximum value. When comparing the behaviour of the box model versions with four models from the CMIP5 archive (Fig. 2.1d), the box models generally project temperatures in the middle of the range. However, there is a substantial spread in the peak warming projected by the simple model. The spread shows the effects relatively subtle changes in the structure of the simple model, specifically in the parametrizations

¹All model versions have been re-optimised independently such that the optimised parameters vary between versions.

of air-sea fluxes, can bring to future projections given the same observational constraints and total emission pathway.

To diagnose the origin of the spread in temperature projections obtained from the simple model versions, we fix each of the five optimised parameters in turn between the models. Fixing the air-sea carbon flux timescale parameter increases the spread of peak warming by 0.05°C . Changes of this order of magnitude are seen when fixing all four parameters except for the air-sea heat flux, which causes an increased spread in peak warming of roughly 1°C . The experiment using this particular idealised framework therefore suggests that in constraining future peak global mean temperature projections under 1 trillion tonne emission mitigation scenarios, the air-sea heat flux is likely the more dominant flux. Varying the carbon flux parametrization does cause a change in the timing of the peak in atmospheric CO_2 concentrations. For example, errors in the modelling of air-sea carbon fluxes can therefore cause ineffective emission mitigation scenarios that aim to limit global warming by 2°C in 2100.

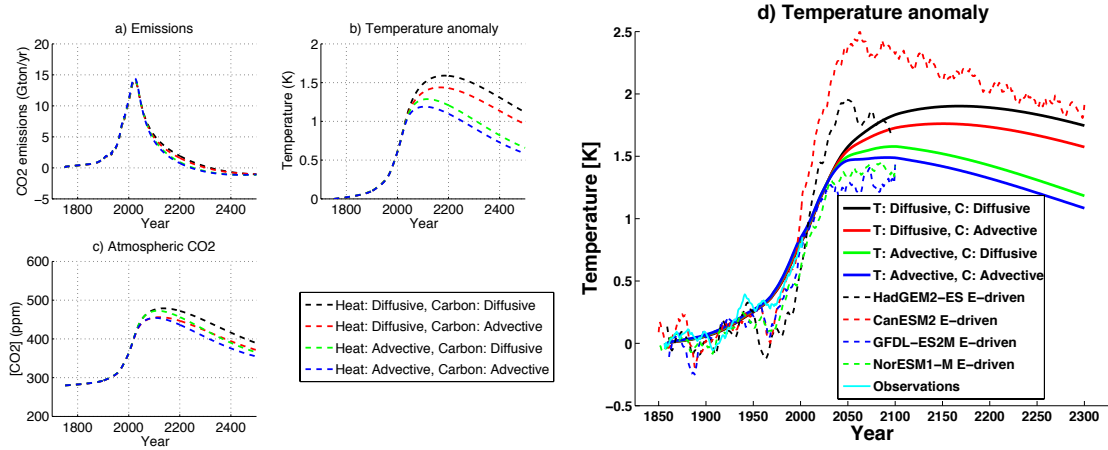


Figure 2.1: a)-c) Model output colour coded according to the structural form of the air-sea parametrizations, where the black curve denotes the original, fully diffusive model from Allen et al. [2009]. The model is driven by historical emissions until 2000 after which the panels show projections. a) CO_2 emissions E after 1750, which includes both anthropogenic and temperature feedback emissions. b) Global mean temperature anomaly above pre-industrial value. c) Atmospheric CO_2 concentrations. d) Temperature anomaly projections of the four simple model versions and four CMIP5 models under scenario RCP2.6 along with observations until 2009 (cyan). The transient climate sensitivity (TCR) of each of the CMIP5 models used are: 2.3°C (CanESM2), 1.3° (GFDL-ESM2M), 1.4°C (NorESM) and 2.5°C (HadGEM2-ES) [Forster et al., 2013].

2.1.2 Ocean timescales in response to forcing

After examining the effects of air-sea heat and carbon uptake parametrizations on global temperature projection uncertainties, we now consider which representation most closely matches the physics of more complex General Circulation Models (GCMs) under a wider variety of scenarios. We compare simple model output with that of HadGEM2-ES, an Earth System model with a full carbon cycle [Collins et al., 2008] (see section 2.2). We recognise that the behaviour of HadGEM2-ES may not match the real world or other CMIP5 GCMs, however, we chose to use just one model to simplify the optimisation. The available HadGEM2-ES experiments are driven by atmospheric CO₂ concentrations therefore we adapt the box models to be driven by the same concentration forcing. Such a modification in the forcing bypasses the carbon cycle component of the simple model leading us to solely focus on differentiating between the air-sea heat flux structures. The two box models used in this section express the air-sea heat flux parametrization through the use of one (advective) or multiple (diffusive) timescales. The goal of these experiments is then to determine which formulation of air-sea heat flux is 1) most appropriate and 2) sufficient compared to HadGEM2-ES.

We examine the response to an artificial forcing profile constructed of a 1% yearly increase in atmospheric CO₂ levels to four times the pre-industrial value (ramp up), followed by a 1% yearly decrease (ramp down). Each leg of the symmetric ramp can be examined separately. The box-diffusion models are initially optimised to the ramp up section only (Fig. 2.2a), then only optimised to the ramp down section (Fig. 2.2b).

When the model versions are optimised to the 1% increase (ramp up), they fail to capture the cooling phase. In both model versions, the global mean temperature anomaly decreases faster than that of the GCM. When the models are optimised to the ramp down however, the diffusive model version captures both the increase and decrease of global mean temperature anomalies. The single timescale advective model however, overestimates the appropriate temperature response as it can only represent the fast climate response.

The key feature of the GCM that is not captured by the box models is the shift in timescales over the experiment: for the ramp up optimised version, the response timescale on the ramp down section is 5 years for both the diffusive and advective temperature

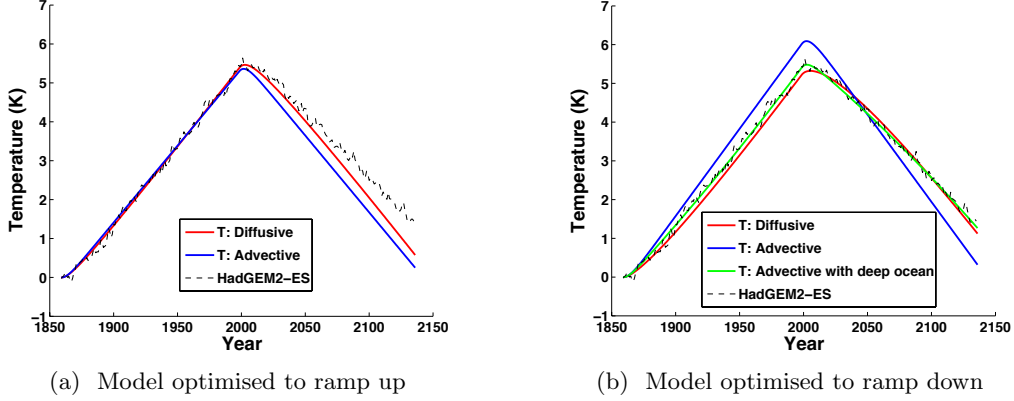


Figure 2.2: Global mean temperature anomalies as a function of time for 1% yearly CO_2 ramp up to quadrupling and ramp down experiment. Solid lines are the box model versions, dashed black line is HadGEM2-ES output. The box model versions are optimised with respect to a) the ramp up section of the HadGEM2-ES output; b) the ramp down section of the HadGEM2-ES output.

versions. For the ramp down, the system adjusts on timescales of 6 and 11 years for the diffusive and advective versions respectively, indicating an increase in the response timescales. The increase in timescale is the result of the system adapting to the forcing and illustrates how a fixed timescale model will fail to behave in a physical manner for future climate scenarios. Changes in ocean dynamics, thermodynamics and biogeochemistry will cause shifts in ocean heat and carbon uptake that a single fixed timescale model will fail to capture. We therefore proceed to introduce a second heat uptake timescale to the advective temperature model. Following Held et al. [2010] and Gregory [2000], we add a deep ocean box with a temperature anomaly T_D and a surface-deep ocean heat exchange H_T proportional to the temperature difference between the upper and lower boxes such that

$$H_T = \zeta(T - T_D), \quad (2.8)$$

where ζ is the inverse of the interaction timescale. To determine whether a 2-timescale advective model with a deep ocean is sufficient to represent the uptake globally, we compare it to the response of the 1% yearly CO_2 increase and decrease when optimised to the ramp down section (Fig. 2.2b). The addition of the second timescale results in a response which follows the output of HadGEM2-ES and outperforms the projection from the purely diffusive model. To explore this further, we use two additional experiments. These consist of a 1% yearly CO_2 increase to $4 \times \text{CO}_2$ followed by (1) a stabilisation; and (2) by an abrupt return back to pre-industrial CO_2 concentrations, shown in Fig. 2.3.

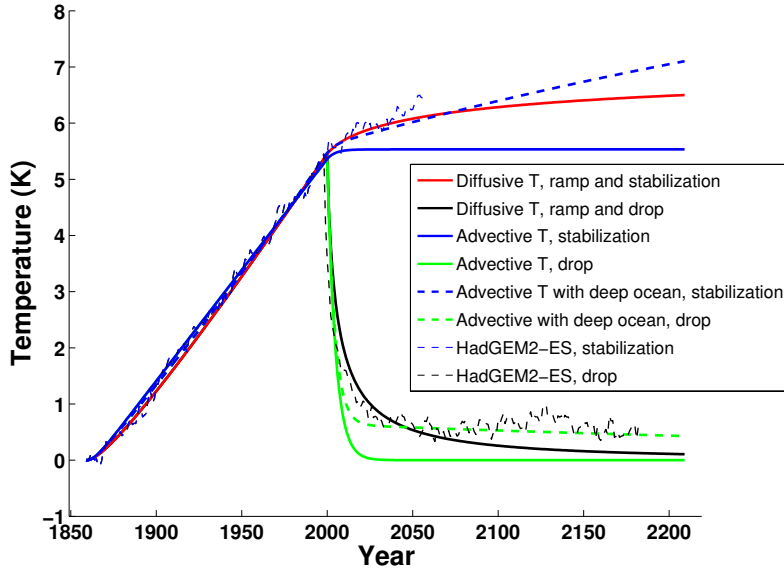


Figure 2.3: Global mean temperature anomalies for HadGEM2-ES and the box models. Up to year 2000, the profile is identical to the 1% yearly CO_2 ramp up for all experiments. Afterwards, the thin blue line shows a stabilisation experiment where CO_2 concentrations are held at four times the pre-industrial value with the HadGEM2-ES model. The same experiment is shown with the solid blue for the single advective timescale model, the dashed blue for the two-timescale advective model and the red line shows the diffusive model. Alternatively, a second set of experiments are performed whereby atmospheric CO_2 levels are abruptly returned to the pre-industrial CO_2 concentration of 280ppm at the year 2000. The solid green curve shows the output of the single advective timescale model for the abrupt return to preindustrial CO_2 , dashed green for the two-timescale advective model, thick black for the multiple timescales diffusive form and the thin black dashed line is the HadGEM2-ES output.

The single advective model again underestimates the response timescales of HadGEM2-ES. The single-timescale advective model results in an equilibrium value of 0°C within 50 years of the abrupt return to pre-industrial concentration. The GCM on the other hand, equilibrates more slowly to a temperature of 0.6°C . The non-zero temperature anomaly in HadGEM2-ES after the drop shows the recalcitrant component [Held et al., 2010] which the advective temperature models fails to capture. The diffusive model, despite its longer adjustment timescale, doesn't capture the recalcitrant component either. The advective model with a deep ocean interaction behaves much more like the GCM and is a vast improvement over the single timescale model. However, the initial adjustment after the step is still much faster than the GCM or the diffusive box model.

We can define a global timescale of ocean carbon or heat uptake by the exponential growth rate of the carbon and heat anomaly stored in the ocean, and compare the simple

	T	C
HadGEM2-ES	200	76
Simple model (diffusive C)		152-155
Simple model (Advective C)		70
Simple model (2-layer ocean)	525	

Table 2.1: Exponential uptake timescales of ocean heat and carbon in HadGEM2-ES and various forms of the simple model.

model timescales to those of HadGEM2-ES. Table 2.1 shows that the simple model’s uptake of heat and carbon is much slower compared to HadGEM2-ES. The discrepancy in the ocean uptake timescales suggests that the response to forcing can be mimicked by simple models only for short-term projections. For longer-term projections, the ocean heat and carbon uptake timescales will change as the climate system adjusts to forcing. To fully understand how future atmospheric CO_2 and temperatures will evolve, a thorough understanding of the time evolution of ocean heat and carbon uptake timescales is required. Throughout this thesis, we will aim to understand how wind stress and surface warming affect the rate and efficiency at which the ocean can absorb atmospheric CO_2 and heat. This in turn will dictate the timescales of how atmospheric global mean temperatures will evolve with time, as shown in section 2.1.1. To quantify the evolution of ocean carbon and heat storage, we will investigate how the ocean carbon cycle responds to temperature and wind stress forcing, which will affect how the ocean absorbs anthropogenic CO_2 emissions. In chapter 5, we will then examine how ocean heat uptake and transport responds to the same forcing.

While simple models such as the 1-D model presented here assume the same form for the air-sea heat and carbon flux, we have shown that the form of the exchange can affect the timing and magnitude of future projections of global mean temperature, even while replicating the same historical trends. We will examine if assuming a similar parametrization for both heat and carbon is sensible by comparing how climate change-induced forcing affects global ocean heat and carbon uptake timescales. The goal is then to firstly calculate climate-carbon feedbacks for atmospheric carbon to understand how ocean carbon timescales may change in the future and secondly, to understand the differences in the ocean heat and carbon uptake and the associated implications for calculating climate-carbon feedbacks and ocean heat uptake.

2.2 HadGEM2-ES

2.2.1 Model description

The atmosphere-ocean coupled model used in this study is HadGEM2-ES (Hadley Centre Global Environment Model version 2 - Earth System version), a primitive equation, general circulation model (GCM) developed by the UK MET office Hadley Centre. The Earth System model is the top-tier version of HadGEM2 and features a full (terrestrial) carbon cycle. The resolution of HadGEM2-ES is the same as HadGEM1, 1.25° of latitude and 1.875° of longitude for an atmosphere with 38 levels, so that grid cells are 208km x 139km at the Equator and 120km x 139km at 55° latitude [METoffice, 2010]. Oceanic horizontal resolution is 1° between the poles and 30° -latitude, transitioning smoothly to $1/3^\circ$ at the Equator [METoffice, 2010], with 40 unevenly spaced vertical levels. The coarse resolution of the ocean model requires parametrizations for eddies following the Gent-McWilliams scheme [Gent and McWilliams, 1990], although the eddy velocity output is not available. All model simulations with HadGEM2-ES discussed in this thesis were performed by the UK MET office.

The addition of the Earth system components from HadGEM1 to HadGEM2 involves inclusion of a full carbon cycle and short lived forcing agents in the form of the HadOCC model. The HadOCC model includes a simple ecological model with one species of phytoplankton and one species of zooplankton, but does not model spatial variations in Redfield ratios [Weber and Deutsch, 2010]. The model achieves a realistic pre-industrial state of climate, vegetation and ocean biology without the need for flux adjustments. The major deficiency of the model is the amplification of errors due to coupling processes such as the under prediction of vegetation in arid regions leading to excessive dust emissions [METoffice, 2010]. More information about the model details can be found in Collins et al. [2008].

Defining the Meridional Overturning Circulation (MOC) index as the maximum in the zonally averaged streamfunction Ψ between 350-1550m depth and 30° - 60° N, the control integration (representing year 1850 conditions) shows a weaker MOC in the Atlantic than estimated from observations. Cunningham et al. [2007] estimate the MOC strength at 18.7 ± 5.6 Sv, whereas HadGEM2-ES's MOC is only 14 Sv (see Fig. 2.4a). Similarly,

the abyssal cell in the model is roughly 6 Sv, while Ganachaud and Wunsch [2000] estimate the abyssal cell strength at 8 ± 2 Sv. Nonetheless, while the strength of the global meridional overturning is weaker, the structure is representative of our understanding of the contemporary circulation.² The patterns of the potential temperature and dissolved inorganic carbon in the Atlantic are also reminiscent of previous observational and model studies, including high levels of DIC in the Southern Ocean and underneath the tropical thermocline (Figs. 2.4b-c, 1.7 and 1.8).

The equilibrium climate sensitivity³ of HadGEM2-ES is 4.59 K, which is near the high end of the CMIP5 models [Andrews et al., 2012]. While the model has a strong climate feedback, HadGEM2-ES is certainly not an outlier within the CMIP5 models (for example, see Fig. 2.1).

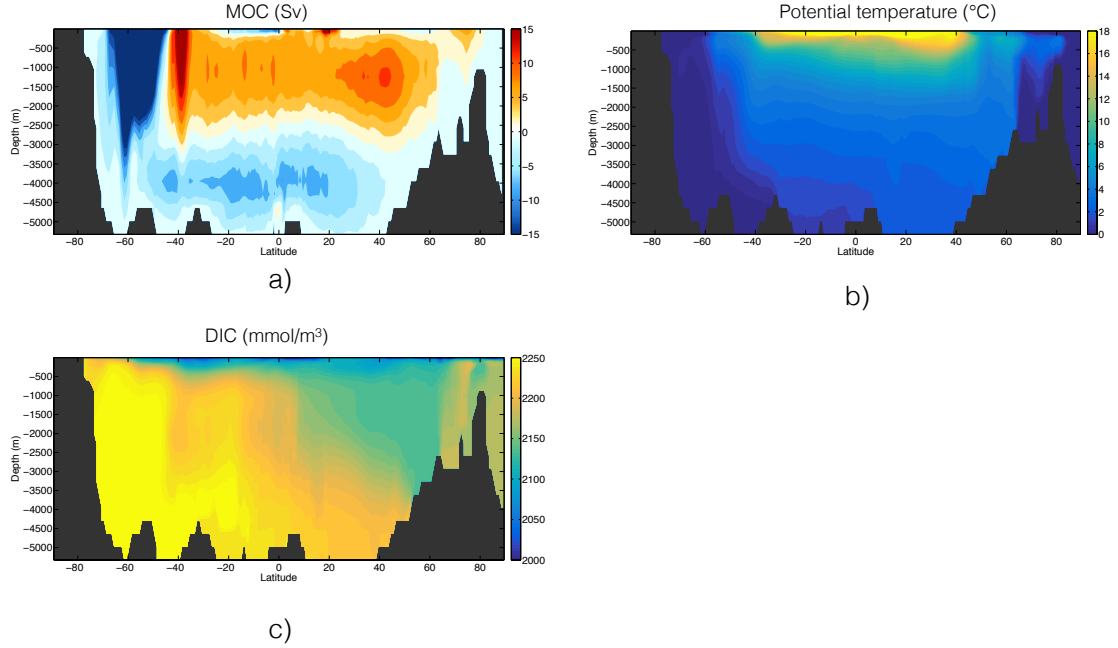


Figure 2.4: HadGEM2-ES Control integration 240-year mean of Atlantic (280-360°E) a) Meridional Overturning Circulation (MOC), b) zonal mean potential temperature (°C) and c) dissolved inorganic carbon (mmol/m³).

²The presence of the strong Deacon cell in the Southern Ocean is the result of the bolus velocities not being made available.

³The equilibrium global mean surface temperature change following a sustained 2 times atmospheric CO₂ perturbation.

2.2.2 Ocean carbon and heat under abrupt atmospheric CO₂ perturbations

The main purpose of using HadGEM2-ES is to investigate the response of the ocean carbon cycle to climate forcing. Particularly, the (quasi-)equilibrium climate response. As part of the IPCC fifth assessment report [Solomon et al., 2007], experiments were performed by the UK MET office whereby the atmospheric CO₂ concentration is instantaneously doubled ($2\times\text{CO}_2$) or quadrupled ($4\times\text{CO}_2$) compared to the pre-industrial level of 280 ppm, and kept constant for 140 years while the coupled climate system partially adjusts to the forcing. After 140 years, the mixed-layer has equilibrated, the upper ocean is close to equilibrium however the abyssal ocean has not reached equilibrium.⁴ To illustrate changes in the ocean circulation occurring as a result of the increased forcing in the step experiments, Fig. 2.5a shows the perturbed overturning streamfunction as the mean of year 130-140 in the $2\times\text{CO}_2$ step experiment. The increase in North Atlantic buoyancy due to temperature and freshwater changes causes a weakening and shallowing of the upper overturning and a weakening of the abyssal cell [e.g. Stommel, 1961, Kuhlbrodt et al., 2007, Schloesser et al., 2012]. Although observations of ocean heat and carbon are transient, the change in Atlantic zonal mean temperature in the HadGEM2-ES $2\times\text{CO}_2$ experiment (Fig. 2.5b) shows similar patterns to the observed warming between 1955-2003 [Levitus et al., 2005] (see Fig. 2.5d), particularly the strong warming at 40°N and 40°S. Similarly, the ocean DIC anomaly in the $2\times\text{CO}_2$ experiment shows similar features to the observed anthropogenic anomaly, including maximum values in the sub-tropical gyres and penetration beneath the thermocline in the North Atlantic (Figs. 2.5b and 1.8b). Similar broad scale patterns are seen in the $4\times\text{CO}_2$ experiments, though the magnitudes and local features differ. The $2\times\text{CO}_2$ and $4\times\text{CO}_2$ experiments are therefore useful tools in assessing quasi-equilibrium changes in the ocean carbon cycle in a fully coupled GCM.

HadGEM2-ES is a fully coupled model, so it includes climate feedback processes between the ocean, atmosphere and the carbon cycle. In this study, we are interested in the ocean carbon cycle response to buoyancy and wind stress forcing. Panel a of Fig. 2.6 shows the zonal mean sea surface temperature (SST) and panel b shows the zonal mean surface wind stress across the Atlantic basin in the control, $2\times\text{CO}_2$ and $4\times\text{CO}_2$ experiments. The temperature profile shows strong warming from the Tropics to 40°S and 40°N. The Northern high latitudes warm more than the Southern high latitudes, showing

⁴Only 140 years are available.

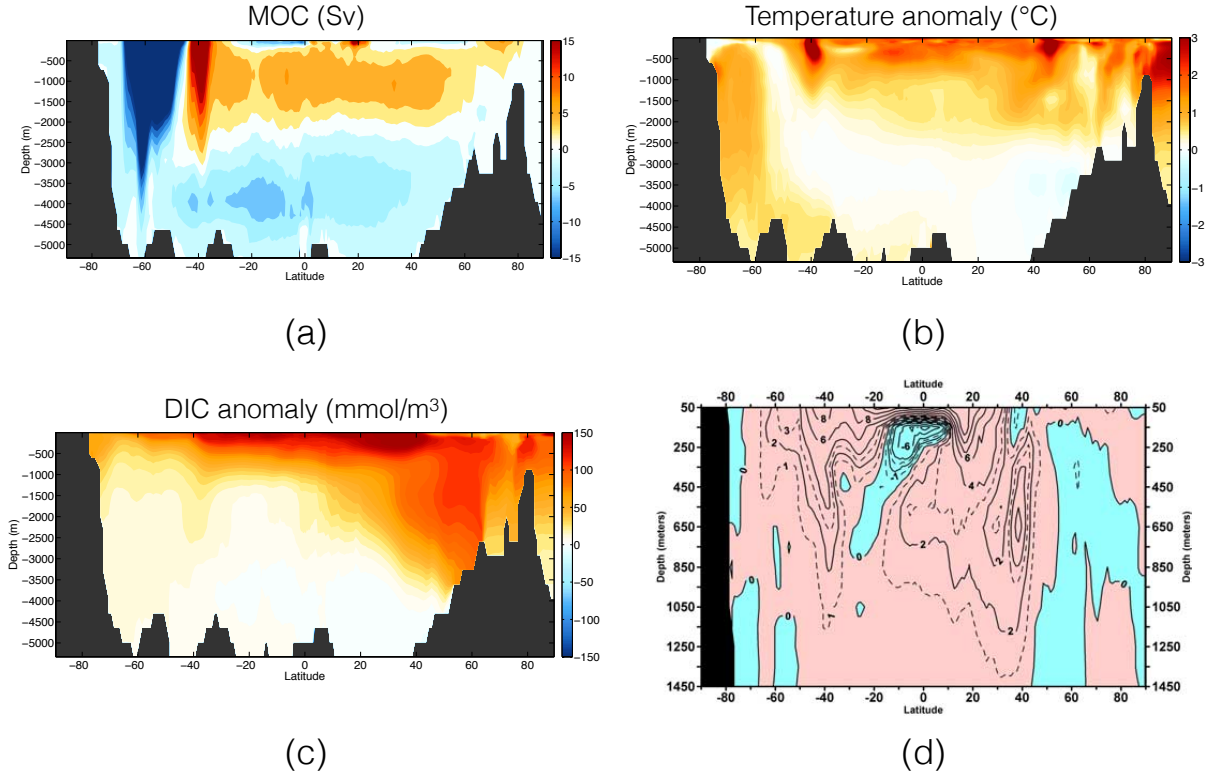


Figure 2.5: HadGEM2-ES Atlantic (280-360°E): a) perturbed Meridional Overturning Circulation (MOC) in the last 10 years of a 140-year long $2\times\text{CO}_2$ experiment; b) Difference in zonal mean potential temperature (°C) and c) difference in dissolved inorganic carbon (mmol/m³) between the last 10 years of the $2\times\text{CO}_2$ experiment and the control run. Panel d shows the observed linear trend in zonal integrated heat content between 1955-2003, with contours spaced $2\times 10^{18}\text{J/yr}$ apart, from [Levitus et al., 2005].

evidence of Arctic polar amplification in the model [Manabe and Stouffer, 1980, Holland and Bitz, 2003]. The Southern Ocean exhibits an increase in surface wind stress (40-60°S) and a relatively weak warming, South of 50°S, since deep mixed layers entrain cold deep waters. As the strength of the surface wind stress over the Southern Ocean increases, the cooling effect increases [Manabe et al., 1990, Xie et al., 2010, Majkut et al., 2014]. In the $2\times\text{CO}_2$ and $4\times\text{CO}_2$ step experiments, there are regions of strong Ekman pumping in the Southern Ocean where sea surface temperatures show local cooling despite the global warming trend.

On the other hand, the surface wind stress over the North Atlantic sub-polar gyre (45-60°N) is reduced in the step experiments. However, the winter-time Northern Annular Mode (NAM), often used as a measure for zonal wind stress strength, is projected to increase in the future climate of CMIP5 ensemble simulations [Karpechko, 2010, Yali and

Huijun, 2010, Gillett and Fyfe, 2013]; although the spread in projections is large and model dependent [Karpechko, 2010]. Nonetheless, models predict changes in the zonal wind in both the Southern and Northern high latitudes as the climate warms. To quantify the ocean carbon cycle response to temperature, salinity and wind stress perturbations, we will use the profiles of zonal mean warming and wind stress changes in HadGEM2-ES, amongst other models and observations, to design a suite of idealized perturbation experiments performed with an ocean general circulation model coupled to a carbon cycle model, described in the next section.

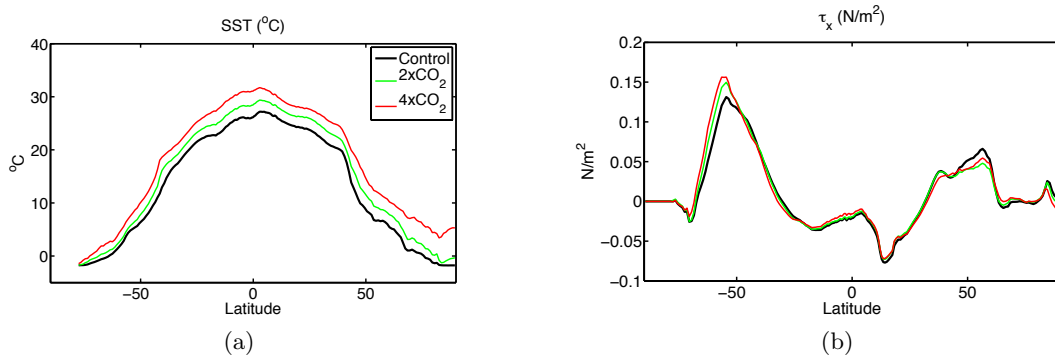


Figure 2.6: Atlantic (280-360°E) zonal mean a) SST (°C) and b) surface wind stress (N/m²) in the control (black), 2×CO₂ (green) and 4×CO₂ (red) experiments.

2.3 MITgcm

2.3.1 Dynamical and Biogeochemical Model

To study the climate-carbon feedbacks of wind stress and temperature changes on atmospheric CO₂, we use an ocean-only idealised experimental set-up which allows us to control the atmospheric forcing without additional feedbacks and to efficiently run a host of perturbation experiments for multiple centuries and even millennia. The model used is the Massachusetts Institute of Technology general circulation model [MITgcm, Marshall et al., 1997]. The MITgcm is a primitive equation model for the ocean and atmosphere based on the incompressible Navier-Stokes equations under the Boussinesq approximation.⁵ The model can be extended using a variety of packages, such as biogeochemical modules, to investigate ocean processes in detail.

⁵In the Boussinesq approximation, density differences in inertial terms are assumed to be small and ignored, so that density differences only appear in gravitational terms [Boussinesq, 1903].

The MITgcm employs a finite volume method for spatial discretisation. Tracer quantities are averaged over the grid cells while the zonal, meridional and vertical velocity components u , v and w respectively are defined at the centre of the orthogonal grid faces (Fig. 7.3). Eddy velocities are not resolved in our simulations but incorporated using the Gent-McWilliams parametrization [Gent and McWilliams, 1990] using the the Griffies [1998] skew-flux formulation for horizontal tracer diffusion. Potential temperature and salinity at the surface is forced with a climatological relaxation similar to Munday et al. [2013]. Such a restoring boundary, as used in other studies [e.g., von der Heydt and Dijkstra, 2008], assumes that the ocean is forced by a steady climatological atmosphere [Haney, 1971]. See section 7.2 for the full equations and variables.

For this study, we include five biogeochemical tracers: dissolved inorganic carbon (labeled C), alkalinity (ALK), phosphate (PO_4), dissolved inorganic phosphorous (DOP) and dissolved oxygen (O_2), using an online biogeochemistry model [Dutkiewicz et al., 2008]. All tracers A follow the general equation:

$$\frac{\partial A}{\partial t} + \mathbf{u} \cdot \nabla A = \nabla \cdot (\boldsymbol{\kappa} \cdot \nabla A) + \mathcal{F}_A, \quad (2.9)$$

with tracer-specific $\mathcal{F}_A$:

$$\begin{aligned} \mathcal{F}_C &= F_{air-sea} + r_{C:P} \mathcal{F}_{PO_4} + J_{Ca}, \\ &= F_{air-sea} + F_{bio}, \end{aligned} \quad (2.10)$$

$$\mathcal{F}_{ALK} = -r_{N:P} \mathcal{F}_{PO_4} + 2 J_{Ca}, \quad (2.11)$$

$$\mathcal{F}_{PO_4} = -\gamma J_{prod} - \frac{\partial F_P}{\partial z} + \lambda_{remin} [DOP], \quad (2.12)$$

$$\mathcal{F}_{DOP} = \gamma J_{prod} - \lambda_{remin} [DOP], \quad (2.13)$$

$$\mathcal{F}_{O_2} = -r_{O:P} \mathcal{F}_{PO_4}, \quad (2.14)$$

for $O_2 > O_{2crit}$, otherwise $\mathcal{F}_{O_2} = 0$. The terms in the equation 2.10-2.14 are:

- $F_{air-sea}$ is the air-sea flux of carbon parametrised following eqn. 1.1 Wanninkhof [1992],
- $r_{C:P}$ is the Carbon to Phosphorous Redfield ratio,

- J_{Ca} is the remineralisation of $CaCO_3$ affecting DIC and ALK in a 1:2 ratio. A constant ratio of $CaCO_3$ remineralisation to organic carbon dissolution of 5% is assumed and the remineralization depth scale is 3.5 km [Sarmiento et al., 2002, Jin et al., 2006],
- $-r_{N:P}$ is the Nitrogen to Phosphorous Redfield ratio,
- The primary production, J_{prod} , is the biological consumption of phosphorous. A simplified parametrization for biological productivity models nutrient (phosphate) and light limitation following Michaelis-Menten kinetics [Dutkiewicz et al., 2008]. A fraction γ of the productivity enters the dissolved organic pool which has a 6 month e-folding timescale. The remaining fraction is exported toward depth [Yamanaka and Tajika, 1996],
- $\frac{\partial F_P}{\partial z}$ is the accumulation of remineralised phosphorous with depth. The downward flux of phosphorous F_P is given by a power law curve: $F_P(z) = F_P(z_c) \left[\frac{z}{z_c} \right]^{-b} = (1 - \gamma) J_{prod} \Delta_z z \left[\frac{z}{z_c} \right]^{-b}$, where z_c is the depth at base of the surface layer of thickness $\Delta_z z$ where production occurs. b is a dimensionless parameter determining the flux attenuation with depth, measured empirically using data from flux traps [Martin et al., 1987] and modelled phosphate distributions [Yamanaka and Tajika, 1996],
- λ_{remin} is the 6-month remineralization timescale of DOP .

See appendix for the values of the constants used (section 7.2). The fractionation of $DIC = [CO_2^*] + [HCO_3^-] + [CO_3^{2-}]$ into dissolved CO_2^* , HCO_3^- and CO_3^{2-} obeys the chemical equations:



The chemical equations proceed to equilibrium in a matter of minutes so we express the relative abundances of the species according the equilibrium constants:

$$K_1 = \frac{[HCO_3^-][H^+]}{[CO_2^*]}, \quad (2.18)$$

$$K_2 = \frac{[CO_3^{2-}][H^+]}{[HCO_3^-]}, \quad (2.19)$$

the values of which depend on temperature, salinity and pressure and are empirically determined [Millero, 1995]. We determine the air-sea flux of carbon, according to the Wanninkhof [1992] parametrization defined above, such that we need the value of ocean $p\text{CO}_2$, $p\text{CO}_2^o$. The ocean $p\text{CO}_2$ is proportional to $[\text{CO}_2^*]$ with a proportionality constant which depends on temperature and salinity. We use the equation for Alkalinity to provide a final constraint so that the system can be solved for $[\text{HCO}_3^-]$, $[\text{CO}_3^{2-}]$, $[\text{CO}_2^*]$, $[\text{H}^+]$ and $p\text{CO}_2^o$ [Follows et al., 2006].

The ocean interacts with the atmosphere via the air-sea flux of carbon such that the well-mixed atmospheric $p\text{CO}_2$, $p\text{CO}_2^a$ is given by the area integral of the air-sea flux:

$$\frac{\partial p\text{CO}_2^a}{\partial t} = \int F_{\text{air-sea}} dA + E = \int K \tau (p\text{CO}_2^a - p\text{CO}_2^o) dA + E, \quad (2.20)$$

where K varies spatially but is constant in time and E is emissions of CO_2 . Usually $K\tau$ is written as $k' u_w^2$, where u_w is the wind speed and k' is a constant. Since $\tau = \rho_{\text{air}} C_d u_w^2$, where ρ_{air} is the density of air and C_d is the drag coefficient, we can re-write $k' u_w^2$ as $K \tau$ where $K \rho_{\text{air}} C_d = k'$. The atmospheric box only contains a well-mixed atmospheric $p\text{CO}_2$ but no temperature or dynamical feedbacks. The specific experimental set-up of the model simulations will be discussed in the following sections.

2.3.2 Experimental set up

The basic model set-up is the same as Munday et al. [2013, 2014] at $2^\circ \times 2^\circ$ horizontal resolution with a modified wind stress forcing profile. The coarse model resolution is chosen such that a large set of perturbation experiments can be efficiently run while still capturing the features of the ocean large-scale circulation. The domain, shown in Fig. 2.7a, is a sector 20° in longitude, extending between 60°S and 60°N with a re-entrant channel between 60°S and 40°S to allow for a circumpolar current. The basin is 5000m deep, except at the channel boundaries, where the domain is extended zonally by 1 grid cell to form a sill of 2500m depth, in order to represent Drake Passage. Such sector model set-ups are used in studies of the Atlantic Meridional Overturning Circulation [Bryan, 1986, Wolfe and Cessi, 2009], parameter sensitivity [Bryan, 1987], ocean stratification [Henning and Vallis, 2004, Wolfe and Cessi, 2010], the impact of ocean circulation on carbon storage [Ito and Follows, 2003, 2005], the Southern Ocean carbon sink [Lovenduski and Ito, 2009] and eddy saturation [Munday et al., 2013].

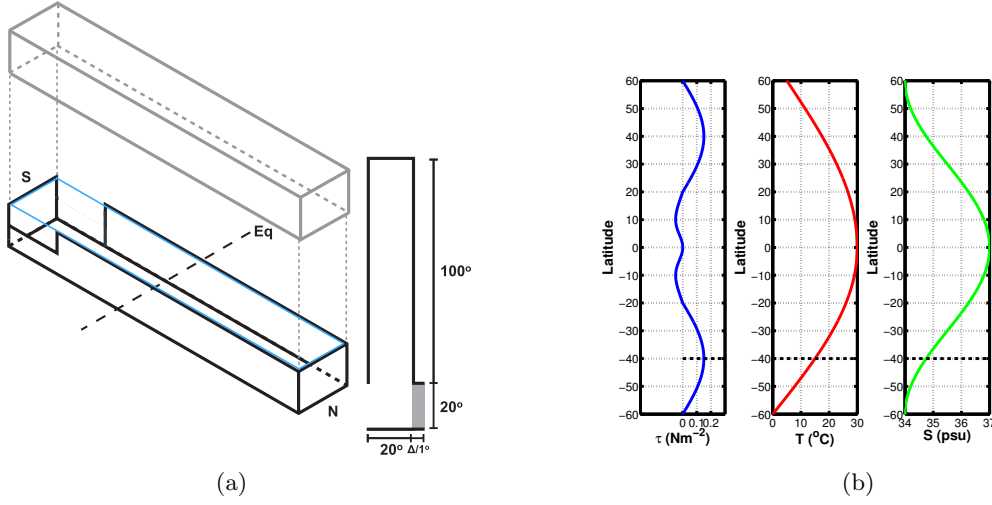


Figure 2.7: MITgcm sector model set-up: a) MITgcm sector geometry, adapted from Munday et al. [2013]. The ocean model depth is 5000 km except in the grey shaded area, one grid cell wide, which is 2500 km deep. The ocean surface is everywhere in contact with an atmospheric box (light grey box) with only a well-mixed $p\text{CO}_2$, b) MITgcm mean state wind forcing (blue), temperature restoring (red) and salinity restoring (green) profiles.

The model has 42 uneven spaced vertical levels, with the surface layer of 10m and the bottom 10 levels of 250m thickness. Grid spacing in the meridional direction is scaled such that the grid boxes are approximately square at all latitudes. The model is forced with a zonal wind stress profile shown in Fig. 2.7b chosen to induce a realistic gyre circulation and equatorial upwelling. The wind stress profile consists of a set of *sin*-functions, modulated at the equator for the appropriate strength of equatorial upwelling:

$$\tau(\Phi) = \begin{cases} \tau_0 \sin[\pi(\Phi + 60)/40], & \text{for } |\Phi| > 20^\circ. \\ 0.6 \cdot \tau_0 \sin[\pi(\Phi + 60)/40] \cdot \left(1 - e^{-(\Phi/8)^2}\right), & \text{for } |\Phi| < 20^\circ, \end{cases} \quad (2.21)$$

with a mean $\tau_0 = 0.15 \text{ N/m}^2$. The wind stress has a seasonal cycle whereby τ_0 oscillates by 0.05 N/m^2 around the mean. Frictional dissipation in the model occurs through viscosity and no-slip boundary conditions. The restoring temperature $\theta^c(\Phi)$ and salinity $S^c(\Phi)$ profiles are chosen such that they broadly match the observed distributions in the Atlantic. The temperature profile is given by the equation:

$$\theta^c(\Phi) = \begin{cases} T_S + \Delta_\Phi T \sin[\pi(\Phi + 60)/120], & \text{for } \Phi < 0. \\ T_N + (\Delta_\Phi T + T_S - T_N) \sin[\pi(\Phi + 60)/120], & \text{for } \Phi > 0, \end{cases} \quad (2.22)$$

where T_S and T_N are the temperature at the Southern and Northern boundary, respectively, and $\Delta_\Phi T$ is the difference in temperature between the Southern boundary and the equator. For the mean state, we use $T_S = 0^\circ$, $T_N = 5^\circ$ and $\Delta_\Phi T = 30^\circ$ with a restoring

timescale of 10 days. The salinity restoring profile is given by:

$$S^c(\Phi) = \begin{cases} S_S + \frac{1}{2}\Delta_\Phi S [1 + \cos(\pi\Phi/240)] & \text{for } \Phi < 0. \\ S_N + \frac{1}{2}(\Delta_\Phi S + S_S - S_N) [1 + \cos(\pi\Phi/240)] & \text{for } \Phi > 0, \end{cases} \quad (2.23)$$

where S_S and S_N are the salinity at the Southern and Northern boundary, and $\Delta_\Phi S$ is the difference in salinity between the Southern boundary and the equator. For the mean state, we use $S_S = 34$ psu, $S_N = 34$ psu and $\Delta_\Phi S = 4$ psu with a restoring timescale of 30 days. The profiles of $T(\Phi)$ and $S(\Phi)$ are shown in Fig. 2.7b. The profiles have been chosen such that the density at the Northern edge lies in between the density at the Southern edge and the Northern edge of the channel, which is required for ensuring an overturning circulation, i.e. there is an overlapping density window between the Southern channel and the Northern edge of the basin [Munday et al., 2013, Wolfe and Cessi, 2010]. The specific variables used are defined in Table 7.1.

2.3.3 Mean state

The wind forcing profile results in the gyre transport shown in Fig. 2.8a. The main features are subtropical gyres North and South of the equator located between 10°-40°N/S with a strength of ± 7 Sv. Ekman transport within the subtropical gyre results in a local depression of the thermocline in the centre of the gyres. In the Northern hemisphere, a sub-polar gyre of strength -6.7 Sv is located between 40°-60°N. Divergence in the centre of the gyre as a result of the wind stress curl causes negative lensing of the thermocline and entrainment of deeper waters into the surface layer. Due to the profile of wind stress, there are also two smaller cyclonic gyres transporting 4 Sv on either side of the equator (between 10°S and 10°N). In the Southern hemisphere, an unbounded barotropic current of 130.2 Sv circles the Southern channel, representative of the Antarctic Circumpolar Current (ACC). Observations put the strength of the ACC between 130-140 Sv [Nowlin and Klinck, 1986, Cunningham et al., 2003].

Figure 2.8 shows the model mean Eulerian (panel b), eddy/bolus (panel c) and residual (panel d, sum of the Eulerian and eddy) overturning stream function. Eddies acts in areas with strong density gradients, so are most active in the Southern channel and the Northern edge of the basin. In the Southern channel, there is a wind-induced Deacon cell of vertically uniform strength 2.2 Sv which extends to -3500m depth (panel a) which is balanced by an eddy-induced cell of opposite sign (panel b) to produce the overturning seen in panel d (the pattern is patchy in the Southern channel due to noise in the eddy

stream function). Figure 2.8e shows an observationally-based estimate of the residual global overturning streamfunction as a function of pressure and latitude. The upper overturning cell is shallower in the model compared to the observations, as in Wolfe and Cessi [2010, 2011], but the overturning streamfunction captures the main features of the meridional and vertical transport: upwelling in the Southern channel, at the equator and in the Northern sub-polar gyre, downwelling in the South and to the North of the Southern channel, in the centre of the sub-tropical gyres and at the Northern boundary.

Similarly, the model has a deep bottom water cell also seen in the observations. The abyssal cell transports -1.3 Sv and the Northern overturning cell, representing the Meridional Overturning Circulation (MOC), has a maximum of 4.2 Sv. Estimates of the abyssal cell are of order -8 ± 2 Sv [Ganachaud and Wunsch, 2000] while recent estimates of the MOC strength are 18.7 ± 5.6 Sv [Cunningham et al., 2007]. Since the idealised model domain is 20° -latitude wide, the model overturning is weaker than observed. If scaled up (assuming the Atlantic is $\approx 3\times$ the width of the idealised basin), the model MOC is only slightly weaker compared to the mean observational estimate.

Near the equator, the model also supports a series of zonal currents. Underneath the mixed layer, there is a wind-driven westward current at the equator. Due to the westward advection, water at the western edge leads to a pressure gradient which drives an eastward jet underneath (Fig. 2.9), similar to the equatorial undercurrent [Knauss, 1960, Philander, 1973, Wyrtki and Kilonsky, 1984, Philander et al., 1987]. The undercurrent is important for delivering cold, nutrient- and carbon-rich water to the eastern side of the basin and contributes to the equatorial outgassing of carbon [Feely et al., 2006] and maintaining zonal sea surface gradients [Bjerknes, 1966]. There is a series of stationary eastward jets on either side of the wind-driven current (Fig. 2.9b), which also advect cold- and nutrient water to the eastern side of the basin. Such an equatorial current system has been observed and well-studied in the past in both the Atlantic and Pacific oceans, and in both observations and models [Tsuchiya, 1972, Firing et al., 1998, Rowe et al., 2000, Izumo et al., 2002, Qiu et al., 2013]. While the structure is similar to observations, the magnitude of the currents in the MITgcm are roughly an order of magnitude smaller, likely due to the model geometry. We also note that the MITgcm shows a strong eastward surface current due to convergence at the western boundary, though it is narrow and shallow.

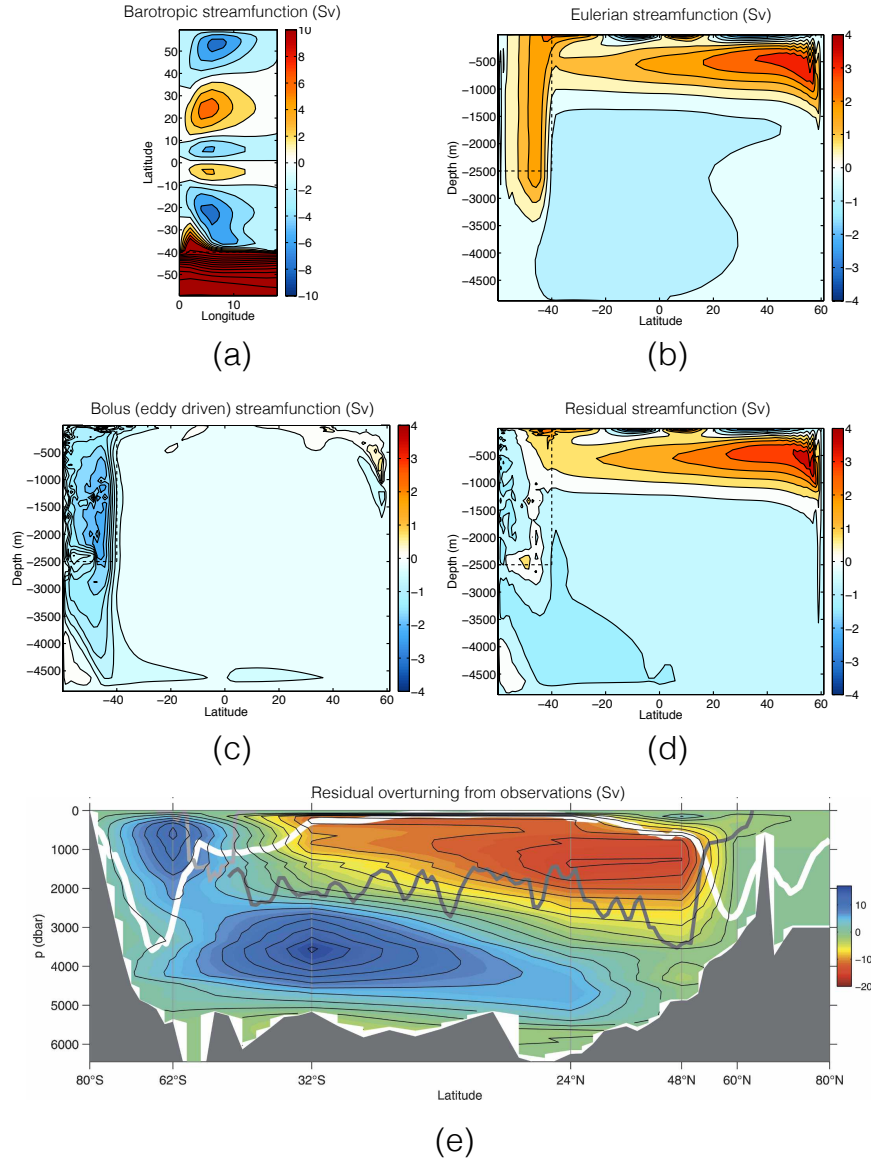


Figure 2.8: Mean state stream functions in the MITgcm sector model (a-d) calculated over the last 100 years of a 20000-year spin up integration; (d) an estimate of the residual overturning streamfunction from observations, from Lumpkin and Speer [2007]. Note that Lumpkin and Speer [2007] use an opposite sign convention.

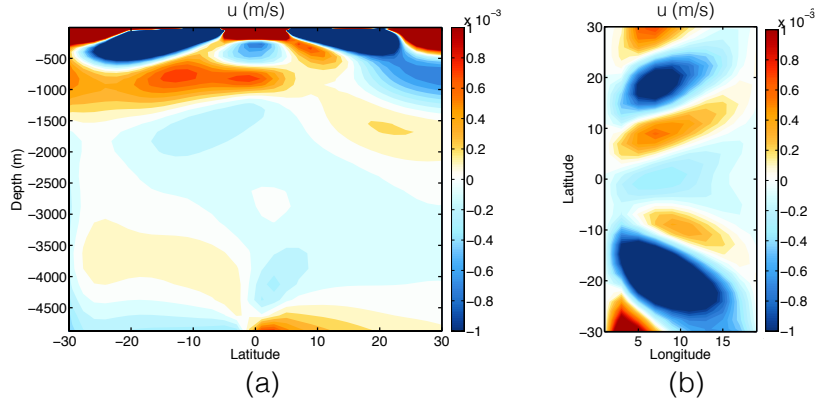


Figure 2.9: a) Zonal mean of the residual zonal velocity u as a function of depth and latitude and b) residual zonal velocity u at a depth of $z=-400\text{m}$ showing the eastward equatorial countercurrents North and South and South the westward equatorial current.

The distribution of temperature and salinity reflects the restoring boundary conditions: temperature and salinity are higher in the Tropics than at high latitudes, with the temperature at the Northern pole higher than the Southern pole (Fig. 2.10a-d). Due to the western boundary currents of the sub-tropical gyres, the tracers show a poleward extension on the western side of the basin. Similarly, the small gyres adjacent to the equator result in convergence on the western side of the basin, at 10°N and 10°S , and divergence at the eastern side of the equator. The divergence at the equator causes local upwelling of colder, fresher water. Circulation and diffusion transports heat and salinity downgradient, so that the surface signals are generally downwelled at the Northern pole and upwelled in the Southern channel. The potential density structure (Fig. 2.10e and 2.10f) is generally temperature-driven: isopycnals are mostly flat outside the channel, except at shallow depth (less than 1000m depth) and North of 40°N . Inside the channel, wind-driven Ekman pumping causes a tilt in isopycnals which is only partially balanced by eddy fluxes. In the mixed layer, the density is also largely reflective of the temperature structure.

Figure 2.11a-c shows the zonally averaged distribution of DIC, alkalinity and phosphate, with higher concentrations at depth. The maximum in concentration is located in the Northern edge of the abyssal cell where the water is oldest. The effect of temperature is also seen by decomposing DIC into its preformed and regenerated components. DIC is decomposed following Gruber et al. [1996]:

$$\begin{aligned} DIC_{reg} &= C_{soft} + C_{carb} = r_{C:P}(P - P_{pre}) + 0.5(Alk - Alk_{pre} + r_{N:P}(P - P_{pre})), \\ DIC_{pre} &= DIC - DIC_{reg}, \end{aligned}$$

where P (phosphate), P_{pre} (preformed phosphate), Alk (alkalinity) and Alk_{pre} (preformed alkalinity) are calculated online and available as model output diagnostics. $r_{C:P}$ and $r_{N:P}$ are the Redfield ratios of carbon to phosphate and nitrogen to phosphate respectively. Fig. 2.11d shows preformed DIC, of which the concentration contours match temperature (there are discrepancies in the top 1500m as a result of C_{dis}). The distribution and magnitude of the biogeochemical tracers agrees well with other studies [e.g., Ito and Follows, 2005] and observations [Feely et al., 2001, Gruber, 1998]. The distribution of regenerated DIC and dissolved oxygen are fairly similar, as oxygen is used up in the remineralisation of organic matter (Fig. 2.11e-f). At the surface, dissolved oxygen is equilibrated with the atmosphere. Waters recently in contact with the surface have high oxygen concentration and are said to be saturated (such as the sinking water at the Southern and Northern boundaries) however those waters have little accumulated regenerated DIC. As the water ages, it is depleted in oxygen and enriched in regenerated DIC, as seen in the abyssal cell. The model shows the presence of an oxygen minimum zone in the mid-latitudes and Tropics between 100-1000m depth, with corresponding high regenerated carbon concentrations. The meridional and vertical extents of the shadow zone compares well with other studies such as Ito and Follows [2005] although the magnitude is slightly higher in our MITgcm model compared to other studies (the minimum in oxygen in the MITgcm model is roughly 170 mol/m^3 , while Ito and Follows [2005] model a minimum between $100\text{-}150 \text{ mol/m}^3$).

The distribution of the biogeochemical tracers is manifested in important mixed layer-only quantities. Convection and upwelling in the Southern channel and convection and wind-induced mixing of nutrients in the Northern sub-polar gyre drive local mixed layer biological activity (Fig. 2.12a). Equatorial upwelling drives biological activity, although the source of equatorial nutrients is Southern Ocean upwelling such that the equatorial waters are already partially depleted in nutrients. As a result, biological activity in the equatorial region is 20–40% of the high latitude value. Such a difference in the magnitudes of the biological activity are supported by satellite chlorophyll observations (NOAA and SeaWiFS). However, in the present model, the equatorial biology is likely too weak and the Southern channel biology is relatively too strong [Williams and Follows, 2011], likely due to a lack of seasonality and light-limitation in the Southern channel (note that the model does not have sea ice). Low productivity is modelled in the oligotrophic⁶ sub-tropical

⁶Environments that have low nutrient and high oxygen concentrations.

gyres, as observed in the real ocean.

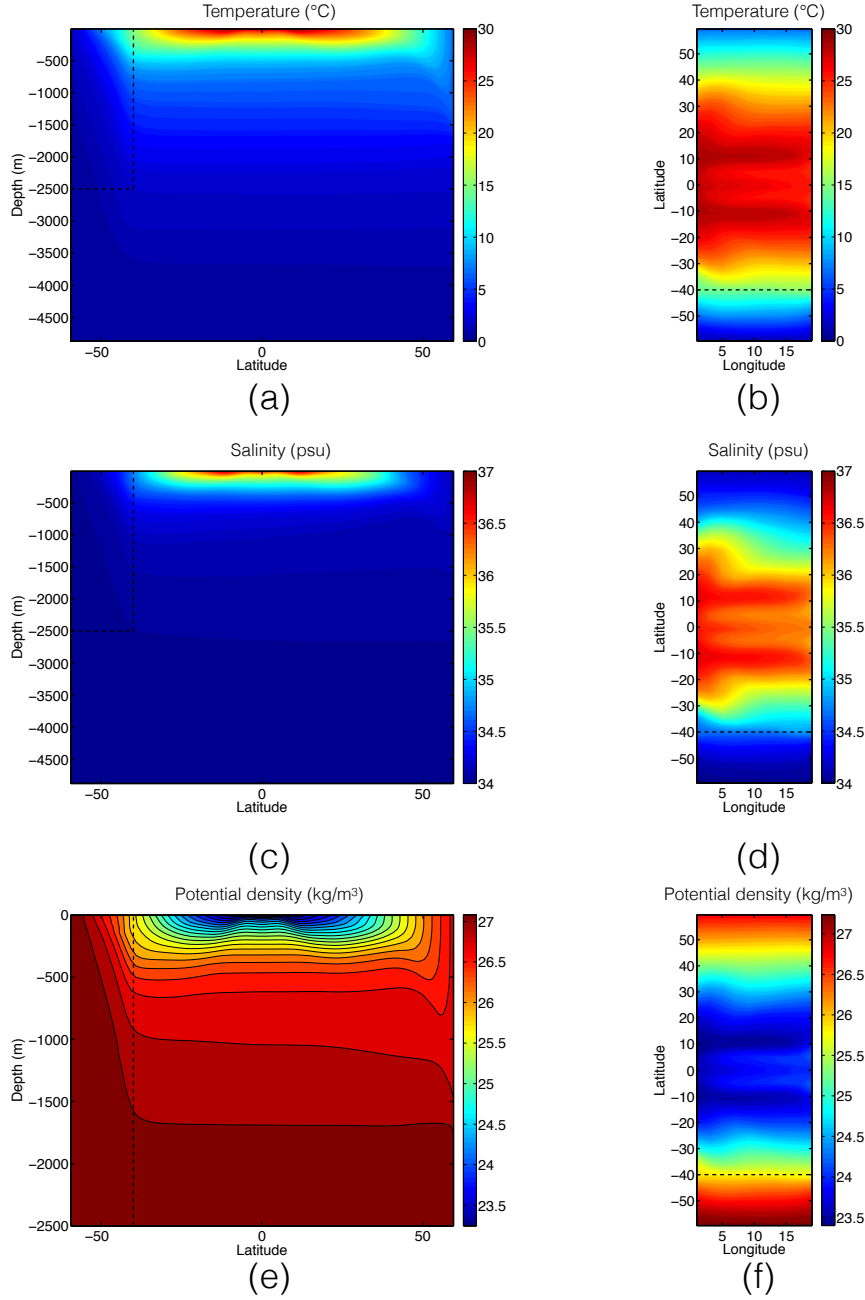


Figure 2.10: Mean state physical quantities in MITgcm sector model: a) Zonally averaged temperature, b) Mixed layer average temperature, c) Zonally averaged salinity, d) Mixed layer average salinity, e) Zonally averaged potential density in the top 1000m and f) Mixed layer average density. The dashed black line marks the extent of the Southern channel.

The distribution of ocean $p\text{CO}_2$ and the air-sea CO_2 flux (Figs. 2.12b,c) provides a complementary view of the biogeochemical and physical processes affecting the tracers' distribution. Upwelling raises the surface DIC and therefore ocean $p\text{CO}_2$ in the Southern channel and the Tropics. In addition, the higher temperatures in the Tropics result in

a further increase of local ocean $p\text{CO}_2$. Cooling of poleward moving waters resulting in increased solubility of CO_2 leads to lower $p\text{CO}_2$ values in the Northern sub-polar region and just North of the Southern channel, particularly in the extension of sub-tropical gyre western boundary currents.

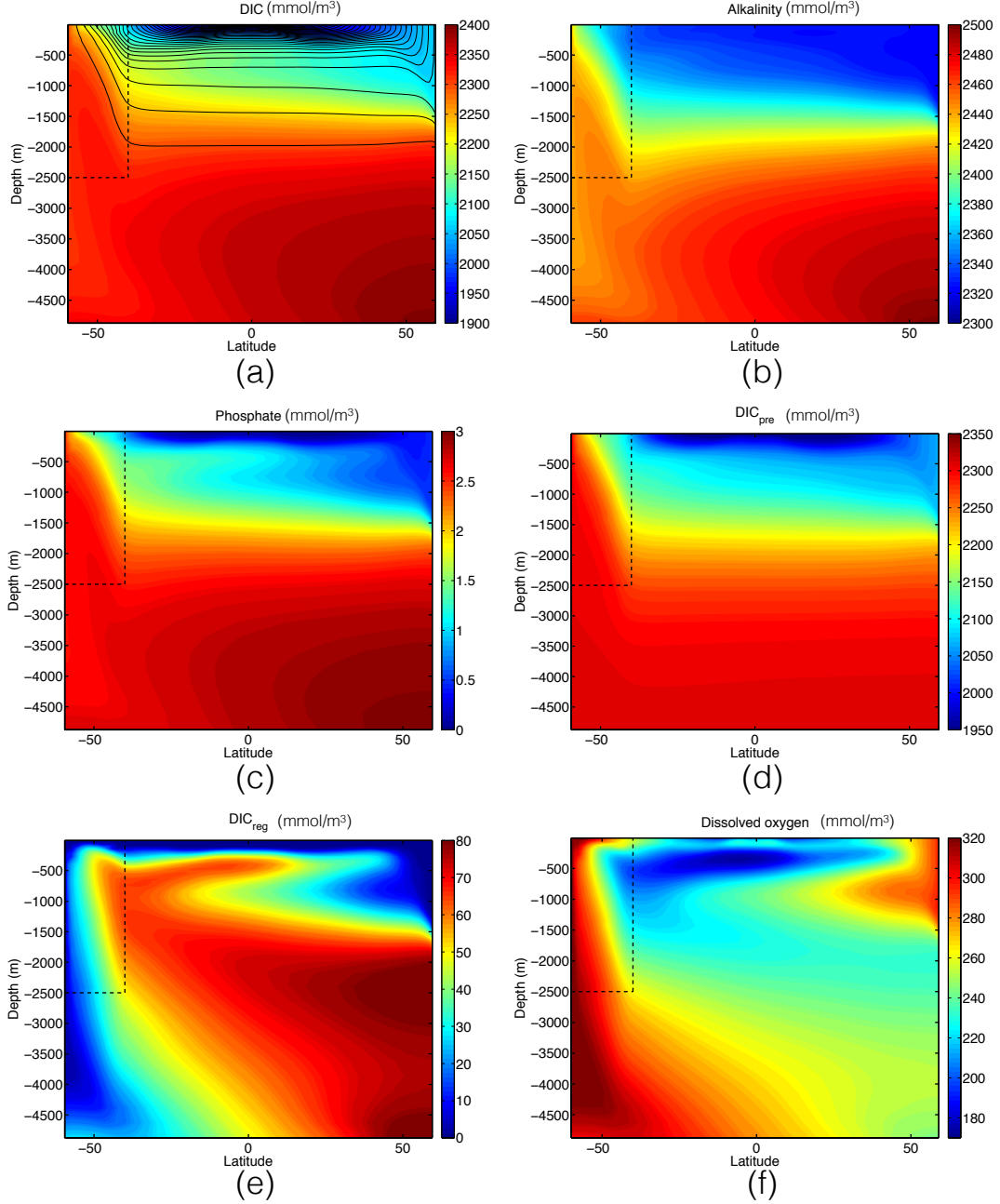


Figure 2.11: Mean state biogeochemical quantities in MITgcm sector model: a) Zonally averaged DIC, where the black lines show surfaces of constant density, b) Zonally averaged Alkalinity, c) Zonally averaged Phosphate PO_4^{3-} , d) Zonally averaged preformed DIC, e) Zonally averaged regenerated DIC and f) Dissolved oxygen.

The magnitude of the ocean $p\text{CO}_2$ is 80 ppm lower than in observations and models

for the period 1980-1995 [Takahashi et al., 2009, Valsala and Maksyutov, 2010] due to the lack of anthropogenic CO_2 forcing in the model. However the spatial pattern and the large-scale meridional gradients of $p\text{CO}_2$ are comparable to observations (on the order of 100 ppm). The mean atmospheric $p\text{CO}_2$ in the model simulation is diagnosed at 302 ppm,⁷ so there is a positive air-sea flux of carbon (i.e. uptake) North of the Southern channel and in the Northern sub-polar region, and a negative air-sea flux of carbon (i.e. outgassing) in the Southern channel and Tropics. The band of positive air-sea CO_2 flux between 30° and 40°S in Fig. 2.12c, where the sub-tropical gyre meets the Southern Ocean, is a good indicator of where intermediate waters are formed. Comparing the pattern with the observations in Fig. 1.6 show good agreement, indicating that intermediate waters are formed in the correct location.

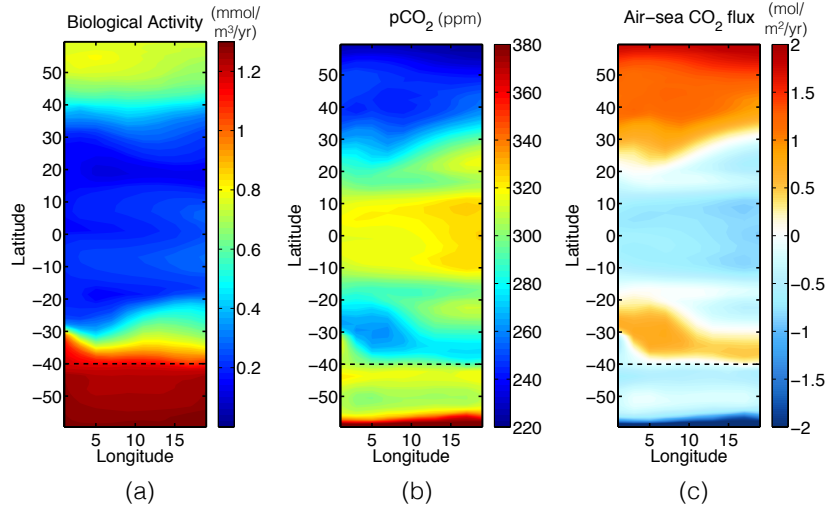


Figure 2.12: Mean state biogeochemical mixed layer quantities in MITgcm sector model: a) Mixed layer average biological activity ($\text{mmol}/\text{m}^3/\text{year}$ of Phosphate), b) Ocean $p\text{CO}_2$ (ppm) and c) Air-sea CO_2 flux ($\text{mol}/\text{m}^2/\text{year}$)

2.3.4 Perturbation experiments

The forcing profiles of zonal wind stress τ , sea surface temperature T and sea surface salinity S lend themselves well for perturbation experiments. In this study, we are interested in the oceanic carbon response to climate change scenarios. Due to the large changes in atmospheric forcing at high latitudes (see Fig. 2.6), we are particularly interested in the ocean response to perturbations in wind stress and buoyancy (temperature and salinity)

⁷The atmospheric $p\text{CO}_2$ in the model is initialised at 280ppm and left to freely evolve during spin-up to reach an equilibrium at 302ppm.

at high-latitudes (e.g., in the Southern Ocean and the North Atlantic). To that end, we design our main experiments as perturbations in:

- Southern Ocean wind stress,
- North Atlantic sub-polar wind stress,
- Global mean sea surface temperature
- North Atlantic sub-polar salinity
- Southern Ocean salinity

The wind stress in both the Southern channel and the Northern sub-polar gyre can be modified in the MITgcm sector model by multiplying the wind stress τ in eqn. 2.21 with a constant over the specific region of interest (see Fig. 2.16a) as in other studies [Lovenduski and Ito, 2009, Wolfe and Cessi, 2010, 2011, Lauderdale et al., 2013, Munday et al., 2013].

Simulating sea surface warming can be done by either adding a constant to the forcing profile of T or multiplying the profile. Across the Atlantic, both observations of sea surface temperatures and simulations using the CMIP3 and CMIP5 ensemble of models show stronger warming in the Tropics than the Southern Ocean [Manabe et al., 1990, Xie et al., 2010, Knutson et al., 2013], pointing towards a multiplicative change in temperature restoring, which is what we choose to adopt. However, we can manipulate the Southern boundary T_S , Equatorial $\Delta_\Phi T$ and Northern boundary T_N temperatures in eqns. 2.22 separately to tune the profile.

The warming in the North Atlantic more ambiguous since a pattern of polar amplification [Manabe and Stouffer, 1980] is seen in observations and models, whereby atmospheric warming is particularly strong in the Arctic due to local feedbacks. However, there is some uncertainty in the high-latitude SST response to climate change since North Atlantic winter mixed layers are deep and act to reduce the surface warming [Manabe et al., 1990, Xie et al., 2010, Knutson et al., 2013]. To mimic the pattern of projected warming, we perturb the ocean surface temperature in the MITgcm sector by multiplying the equatorial warming $\Delta_\Phi T$ in eqn. 2.22 by a factor β . T_N is multiplied by a factor $2\beta - 1$ which ensures that the percentage increase is twice that of $\Delta_\Phi T$, i.e. if $\Delta_\Phi T$ is increased by 10%, T_N is increased by 20%. For these experiments which we will call our “base” warming

scenario, $T_S = 0$. Such perturbations will ensure warming in the Northern sub-polar gyre and Southern ocean, along with a decrease in the shared density window between the two poles. In addition, sensitivity studies are run to understand how the model response changes due to polar warming, which is simulated by varying the percentage change of T_N and T_S relative to $\Delta_\Phi T$. These are discussed in details in the relevant section (section 3.1.4, Fig. 5.9).

We will also estimate the impact of high latitude buoyancy forcing without any warming perturbations affecting the solubility of carbon by perturbing the salinity restoring profiles. We perturb the salinity restoring parameters S_N at the Northern pole and S_S at the Southern pole, in eqn. 2.23. Such perturbation scenarios mimic the buoyancy effects similar to the “warm pole” experiments by Wolfe and Cessi [2010] without the associated warming effects. Figure 2.13 shows a set of example perturbed forcing profiles.

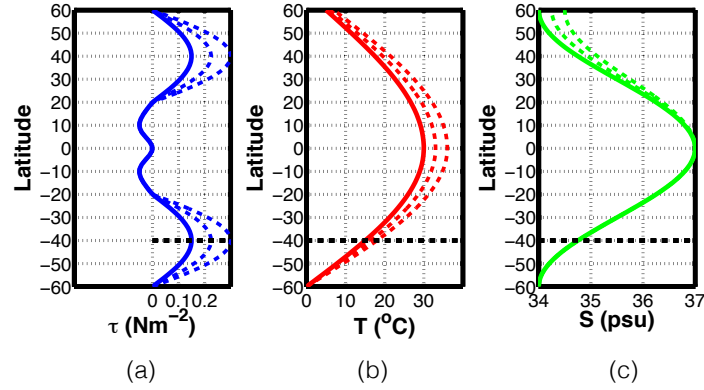


Figure 2.13: MITgcm sector model forcing profiles (solid lines) with example perturbed profiles (dashed lines) for a) zonal wind stress, b) temperature and c) salinity. The wind stress perturbation plot shows $1.5\times$ and $2\times$ Southern Ocean and North Atlantic wind stress profiles. The temperature perturbation shows a $1.1 \times T$ and a $1.2 \times T$ profile. The salinity perturbation shows a $S_N + 0.25$ and a $S_N + 0.5$ psu profiles.

To summarise, the following table lists the set of perturbation experiments run along with the symbol that will be used to refer to the particular perturbation:

Perturbation	Symbol
Multiplication of zonal wind stress τ between 60° and 20°S	τ_{SO}
Multiplication of zonal wind stress τ between 20° and 60°N	τ_{NA}
Multiplication of the pole-to-equator temperature restoring difference ΔT by β and the Northern boundary restoring temperature T_N by $2\beta-1$	T
Reduction of the Northern boundary salinity restoring S_N	S_N
Reduction of the Southern boundary salinity restoring S_S	S_S

The purpose of this thesis is to quantify climate-carbon feedbacks under different forcing scenarios. To that end, the MITgcm perturbation experiments serve to examine relationships for predicting the sensitivity of atmospheric CO₂ to wind stress and temperature buoyancy forcing in a complex system and identify additional feedbacks and non-linearities. The next section describes the basic framework used throughout the thesis to calculate the atmospheric $p\text{CO}_2$ response to forcing.

2.4 Mixed layer carbon feedback

2.4.1 Mixed layer carbon flux equation

The aim of this study is to quantify the impact of wind stress and temperature (both buoyancy and solubility effects) on atmospheric CO₂. A knowledge of the mixed layer carbon concentration is therefore required, which determines the magnitude of the carbon flux in or out of the ocean. We therefore derive a general equation for mixed layer carbon concentration. Later on, the equation will be simplified for each region considered depending on the dominant balance of local processes.

We begin with the general prognostic equation for a tracer C , dissolved inorganic carbon:

$$\frac{\partial C}{\partial t} = -\nabla \cdot [(\bar{\mathbf{u}} + \mathbf{u}^*)C] + \nabla \cdot (\boldsymbol{\kappa} \cdot \nabla C) + F_{air-sea} + F_{bio} \quad (2.24)$$

where $\bar{\mathbf{u}}$ is the Eulerian velocity, $\mathbf{u}^*$ is the bolus velocity and $\bar{\mathbf{u}} + \mathbf{u}^*$ is the residual velocity [Marshall and Radko, 2003], $\boldsymbol{\kappa}$ is the diffusivity matrix, $F_{air-sea}$ is the air-sea flux and F_{bio} the biological flux, as defined in section 2.3.1. We then integrate from the bottom of the mixed layer $z = -H$ to the free surface $z = \eta$ and define the mixed layer thickness $h = H + \eta$ and the mixed layer carbon concentration $C_h = \frac{1}{h} \int_{-H}^{\eta} C dz$. In doing so we assume the following:

- Carbon is well-mixed within the mixed-layer, $\frac{\partial C_h}{\partial z} \approx 0$, except at $z = -H$ where $C(-H) = C_h$.
- The only sources and sinks in the mixed layer are the biological flux, F_{bio} , and the air-sea flux, $F_{air-sea}$, the latter only acting at $z = \eta$.

For the details of the integration over the mixed-layer, see appendix section 7.3. The integrated equation is given by:

$$\begin{aligned}
h \frac{\partial C_h}{\partial t} = & - C_h(w_h + u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y} + \frac{\partial h}{\partial t}) + \frac{\partial C}{\partial x} u h + \frac{\partial C}{\partial y} v h \\
& + \left[\kappa_x \frac{\partial^2 C_h}{\partial x^2} + \kappa_y \frac{\partial^2 C_h}{\partial y^2} \right]_h + \left(\kappa_z \frac{\partial C_h}{\partial z} \right)_{z=-H} \\
& + F_{air-sea} + [F_{bio}]_h,
\end{aligned} \tag{2.25}$$

where $w_h = w(-H)$ and we define $[...]_h = \int_{z=-H}^{z=\eta} ... dz$. To understand the magnitude and patterns of the terms in equation 2.25, we calculate the 240-year means of the terms in the HadGEM2-ES pre-industrial control run⁸ (Fig. 2.14) and in the MITgcm pre-industrial control integration (Fig. 2.15).

The term $C_h w_h$ is the vertical advection of carbon at the bottom of the mixed layer (panel a in Figs. 2.14 and 2.15) and $C_h \left(u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y} \right)$ is the contribution from horizontal advection of carbon across a sloping mixed layer (panels b and c). The term $C_h \frac{\partial h}{\partial t}$ measures instantaneous mixing but is only relevant for un-equilibrated systems and won't be considered in the present work. Owing to weak horizontal gradients in C , the terms $\frac{\partial C}{\partial x} u h$ and $\frac{\partial C}{\partial y} v h$ are small compared to the other terms in the equation as shown in panels d and e, so these terms will henceforth be neglected (in the Southern Ocean, their magnitude is larger than elsewhere, but still an order of magnitude smaller than the other advection terms). Due to the weak horizontal carbon gradients, the horizontal diffusion within the mixed layer is smaller than the advection terms by two orders of magnitude (panel g). The contribution of the retained advective terms (panel f) is shown to be significant over all regions of the ocean.

The biological flux (panel h) is always negative in the mixed layer and is of similar magnitude to the air-sea flux (panel i), apart from the Northern sub-polar regions. Compared to biology, the air-sea flux and advection terms (panels a and f), diapycnal diffusion across the mixed layer base (panel g) is negligibly small. Biology and the air-sea flux are also comparable to the advection terms, though the strength of the biology is relatively weaker, but comparable, in the MITgcm. Levy et al. [2013] suggests biology and the air-sea flux are of similar magnitude, so it is likely that the biology flux is slightly too weak in the MITgcm, the caveats of which are discussed in the relevant chapters. The patterns

⁸As mentioned previously, the calculations do not include the eddy velocities u^* and v^* as they are not output by the model.

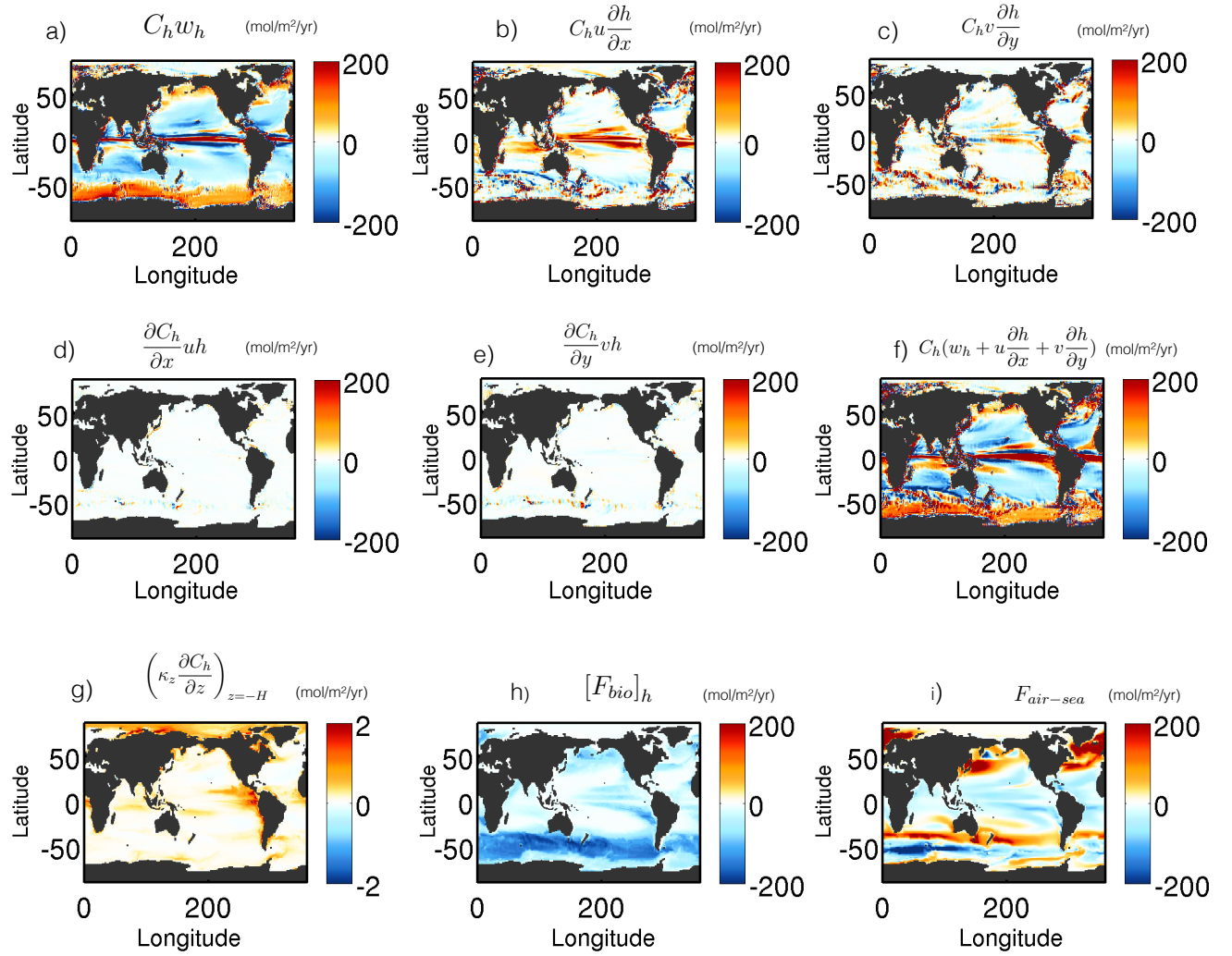


Figure 2.14: Mixed layer carbon terms from eqn. 2.25 diagnosed in the 240-year average of HadGEM2-ES control run ($\text{mol/m}^2/\text{year}$).

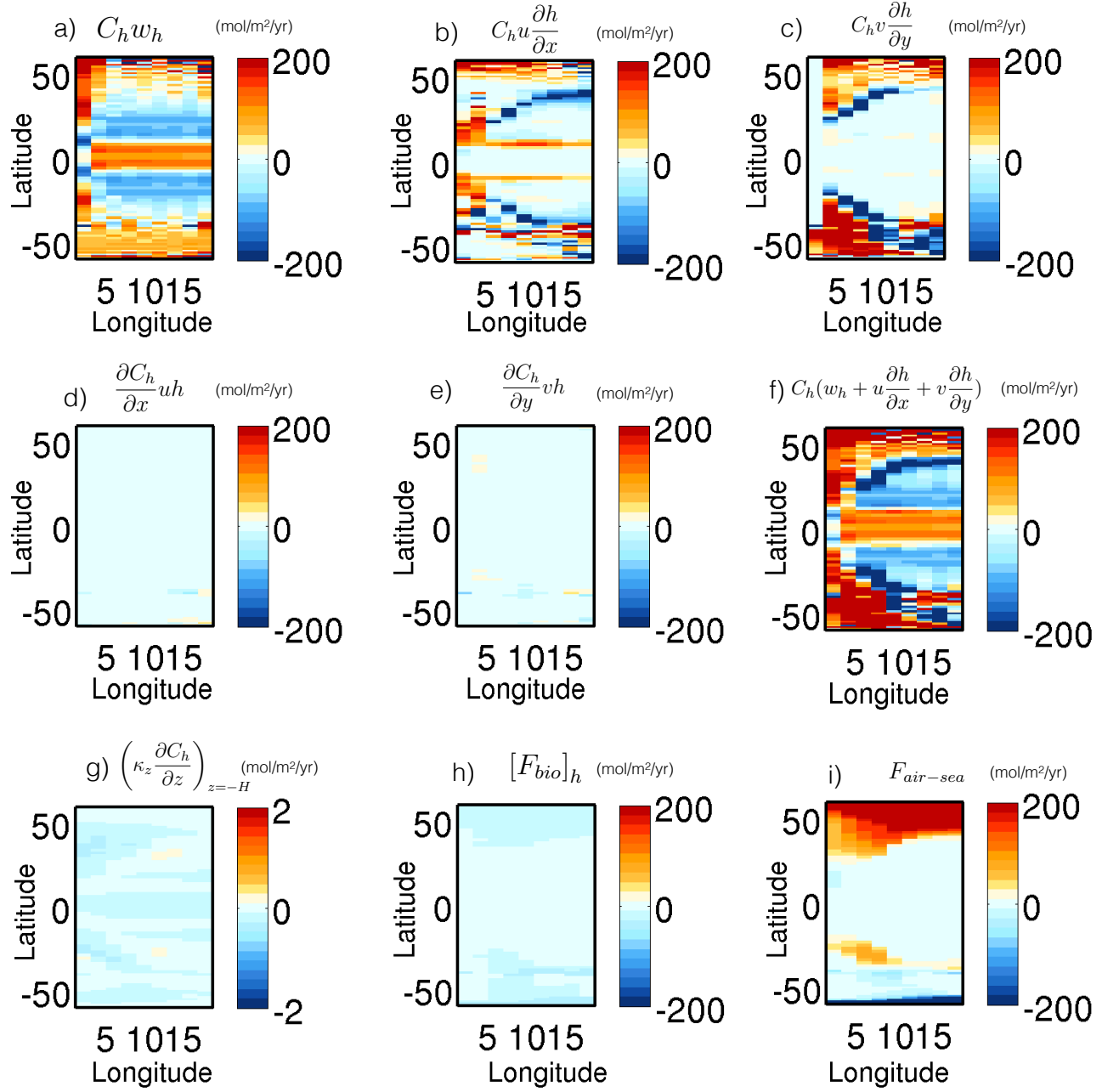


Figure 2.15: Mixed layer carbon terms from eqn. 2.25 diagnosed in the MITgcm control run (mol/m²/year).

and magnitudes of mixed-layer carbon processes calculated here are otherwise similar to Levy et al. [2013] who use observational data sets. We therefore preserve the leading terms in the mixed-layer carbon budget: advection, biology and air-sea flux terms leading to

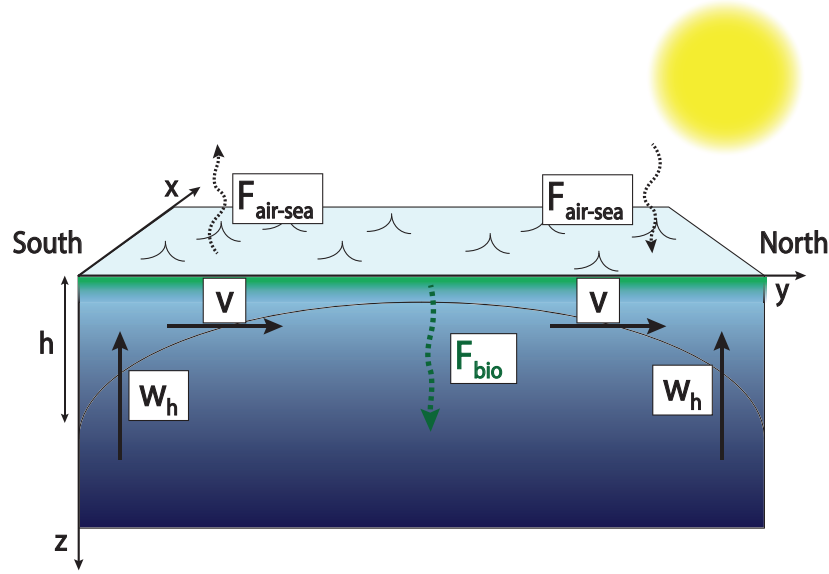
$$h \frac{\partial C_h}{\partial t} = -C_h(w_h + u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y}) + F_{air-sea} + [F_{bio}]_h. \quad (2.26)$$

To illustrate the physical meaning of each term, Fig. 2.16a shows a diagram of the mixed layer fluxes in an ocean basin with a sloping mixed layer vertical extent h in the y -direction. In addition, table 2.16b lists the timescales associated with each process in both the Southern and Northern high latitudes. The timescales are a relative measure of how quickly a certain process can affect the mixed layer carbon concentration. The timescales are calculated using typical values of the parameters in the Northern and Southern high latitudes from the HadGEM2-ES and MITgcm control integrations. The left hand side of equation 2.26 is set equal to $h \frac{\partial C_h}{\partial t} \approx h \frac{C_h}{t_0}$ and then the equation is solved for the timescale t_0 . For every term on the right hand side, we can then estimate the order of magnitude of t_0 for each process. While the advective fluxes are comparable to the air-sea and biological flux in terms of magnitude, the timescales of adjustment are on the order of decades (as a result of the dependency of the advective flux on C_h).

Equation 2.26 is the general mixed layer carbon flux equation and will be the starting point for the calculations in the remainder of the thesis. The equation can be used to calculate the change in C_h given a change in any forcing f as long we know how each of terms in equation 2.26 varies with f . We are interested in the equilibrium solution such that $\frac{\partial C_h}{\partial t} = 0$. However, to calculate the response in the air-sea flux we need to know the response in the ocean $p\text{CO}_2$. Calculating ocean $p\text{CO}_2$ can be accomplished by expanding $F_{air-sea}$ in terms of the difference in $p\text{CO}_2$ of the atmosphere, $p\text{CO}_2^a$, and ocean, $p\text{CO}_2^o$, as defined in eqn. 1.1, and solving for $p\text{CO}_2^o$. Once we have an expression for $p\text{CO}_2^o$ using eqn 2.26, we can use the expression to calculate the resulting atmospheric $p\text{CO}_2$ change, i.e. we can calculate the climate-carbon feedback.

2.4.2 Equilibrium feedback equation

In this section, a general equation to estimate atmospheric carbon feedbacks is derived. The derivation presented is a process-based approach which differs from previous studies [Hansen et al., 1984, Friedlingstein et al., 2006, Goodwin et al., 2009] since the aim is to decompose the global climate feedback with respect to a given change in forcing (such as wind stress or buoyancy) in a specific region.



(a) Diagram of mixed layer fluxes in a mixed layer of depth $h(y)$. Note that the horizontal advection term only acts in the presence of a slope in the mixed layer depth.

Timescale τ_0 (years)	North (HadGEM2-ES / MITgcm)	South (HadGEM2-ES / MITgcm)
Vertical advection	15 / 10	5 / 5
Horizontal advection	40 / 10	50 / 10
Vertical diffusion	>10,000 / >10,000	1500 / >10,000
Air-sea flux	<1 / <1	<1 / <1
Biological flux	<1 / <1	<1 / <1

(b) Table of timescales. Values are calculated using averaged quantities between $40^\circ - 60^\circ\text{N}$ and $60^\circ - 40^\circ\text{S}$.

Figure 2.16

The studies by Goodwin and co-authors [Goodwin et al., 2007, 2008, 2009, Goodwin and Lenton, 2009, Goodwin and Ridgwell, 2010] look at global net changes in the individual reservoirs of ocean carbon, which is certainly important for future climate change studies. Similarly, Gregory et al. [2009] calculate the net global climate-carbon feedback in future climate change scenarios. Both Gregory et al. [2009] and the studies by Goodwin and co-authors therefore focus on the global climate feedback without detailed study of the importance of different regions, nor the importance of the physical processes and parameters that determine the overall strengths of the feedback. The work presented here aims to determine the importance of local changes in the Southern Ocean and North Atlantic on the global climate-carbon feedback. We focus on the physical processes involved in de-

termining the relationship between atmospheric CO_2 and the physical, thermodynamical and biogeochemical forcings in the system, as well as the physical parameters that determine the overall strength of the feedbacks. We therefore provide a more in-depth physical understanding of the regional contribution of the Southern Ocean and North Atlantic to global carbon feedbacks. The forcings studied, wind stress, temperature and salinity, are chosen such that they are driven by anthropogenic climate change, establishing a link to the carbon-climate feedback.

The value of $p\text{CO}_2^o$ as a function of a forcing $\mathbf{f}$, $p\text{CO}_2^o(\mathbf{f})$ is obtained by using the equation for global air-sea carbon flux in the absence of anthropogenic carbon emission (eqn. 1.1). At equilibrium, the globally area integrated flux over the whole ocean vanishes such that:

$$\int F_{\text{air-sea}} dA = \int K\tau(p\text{CO}_2^a - p\text{CO}_2^o(\mathbf{f})) dA = 0, \quad (2.27)$$

where we average over the seasonal cycle. In this study we are interested in the high latitude $p\text{CO}_2^o$ response to local forcing so we split the integral into a high latitude area A_H where the forcing $\mathbf{f}$ is applied and the remaining area in the ocean, A_E (see Fig. 2.17):

$$A_H \overline{\tau_H p\text{CO}_{2H}^o(\mathbf{f})} + A_E \overline{\tau_E p\text{CO}_{2E}^o} - (A_H \overline{\tau_H} + A_E \overline{\tau_E}) p\text{CO}_2^a = 0, \quad (2.28)$$

where the overline indicates an area-mean and the subscript ‘ H ’ refers to the high latitude region and ‘ E ’ refers to the remainder of the ocean. Equation 2.28 can be solved for $p\text{CO}_2^a$ to give $p\text{CO}_2^a$ in terms of the high latitude ocean $p\text{CO}_{2H}^o$. However, we note that in the mixed layer carbon analysis in section 2.4.1, $\overline{p\text{CO}_{2H}^o(\mathbf{f})}$ is solved for, rather than $\overline{\tau_H p\text{CO}_{2H}^o(\mathbf{f})}$. We approximate $\overline{\tau_H p\text{CO}_{2H}^o(\mathbf{f})}$ and $\overline{\tau_E p\text{CO}_{2E}^o}$ by $\overline{\tau_H} \overline{p\text{CO}_{2H}^o(\mathbf{f})}$ and $\overline{\tau_E} \overline{p\text{CO}_{2E}^o}$, respectively and thereby introduce a constant error smaller than 3% in the models and regions used in this study.⁹ The feedback equation is then given by:

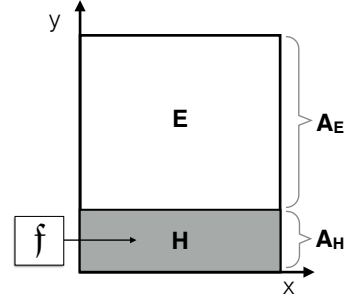


Figure 2.17: For a rectangular ocean basin shown, we label the high latitude region H where we calculate the local response in $p\text{CO}_2^o$ to a forcing $\mathbf{f}$, and a region E where no forcing is applied.

$$p\text{CO}_2^a = P_H \overline{p\text{CO}_{2H}^o(\mathbf{f})} + P_E \overline{p\text{CO}_{2E}^o}, \quad (2.29)$$

⁹The exact error depends on the details of the area split, but remains constant with wind stress so the response to forcing is unaffected.

where $P_H = \frac{A_H \overline{\tau_H}}{A_H \overline{\tau_H} + A_E \overline{\tau_E}}$ and $P_E = \frac{A_E \overline{\tau_E}}{A_H \overline{\tau_H} + A_E \overline{\tau_E}}$. The proportionality term P_H can be larger than A_H in areas of high wind stress such as the Southern Ocean, resulting in a strong influence of $\overline{pCO_{2H}^o(f)}$ on pCO_2^a . Since we are interested in atmospheric pCO_2 changes in response to a forcing f , we express eqn. 2.29 in terms of perturbations to the equilibrium unperturbed state.¹⁰:

$$\Delta pCO_2^a = \Delta P_H \overline{pCO_{2H}^o} + P_H \Delta \overline{pCO_{2H}^o(f)} + P_E \Delta \overline{pCO_{2E}^o(f)} + \Delta P_E \overline{pCO_{2E}^o}. \quad (2.30)$$

Equation 2.30 therefore allows us to calculate the change in pCO_2^a due to a change in $\Delta \overline{pCO_{2H}^o(f)}$ as the ocean adapts to changes in f . High latitude areas are connected to the rest of the ocean via dynamical and biogeochemical processes, so a local change in forcing f induces a local change $\Delta \overline{pCO_{2H}^o(f)}$ and consequentially a non-local change $\Delta \overline{pCO_{2E}^o(f)}$.

To account for the non-local effect of the forcing f on $\overline{pCO_{2E}^o}$, consider the box model shown in Fig. 2.18. The model is similar to that of Toggweiler [1999] and will be used in later chapters to interpret some of the results. The model consists of an atmosphere of carbon concentration C_a and mass M_a that interacts with the surface ocean boxes in the Southern and Northern hemispheres with carbon concentrations C_S and C_N , and masses M_S and M_N through air-sea fluxes F_1 and F_2 , respectively. Representing the overturning circulation, the volume transport Ψ connects the carbon concentrations C_S , C_N and a deep ocean with carbon concentration C_D . The carbon concentration in the South C_S is further enriched through the overturning Ψ and a vertical mixing term. The mixing term is proportional to the difference $C_D - C_S$ and depends on the strength of the mixing f . Our work as a whole is concerned with the effect of a forcing f at high latitudes, so for now let the forcing f be applied to the Southern high latitude ocean. The resulting change in C_N is therefore of interest. The change in C_S due to the forcing f is referred to as the local feedback, whereas the change in C_N due to f is the non-local feedback.

¹⁰Where we define the independent variables to be P_H , P_E , $\overline{pCO_{2H}^o}$ and $\overline{pCO_{2E}^o}$

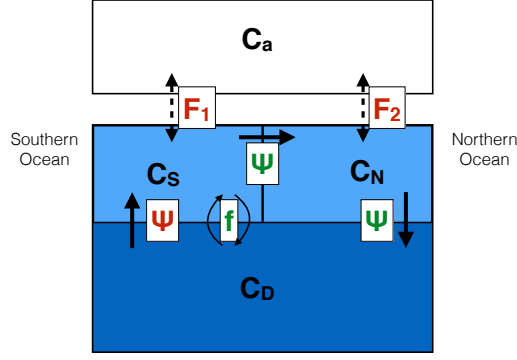


Figure 2.18: Simple box model adapted from Toggweiler [1999] with surface ocean boxes with carbon concentrations C_S and C_N over a deep ocean box with C_D and an atmosphere with C_a . F_1 and F_2 represent air-sea fluxes, Ψ is the overturning circulation and f represents vertical mixing.

The steady state transport of carbon in and out of the deep box is then given by

$$\Psi C_N - \Psi C_D - f [C_D - C_S] = 0, \quad (2.31)$$

such that the deep carbon concentration in steady state is:

$$C_D = (\Psi C_N + f C_S) (\Psi + f)^{-1}. \quad (2.32)$$

The upper ocean carbon concentration in the Northern box can be obtained by combining the previous expression with conservation of mass in the absence of carbon emissions, i.e. $M_A C_a + M_N C_N + M_S C_S + M_D C_D = I$, where I is the total carbon inventory, leading to

$$C_N = \left[\frac{I}{M_D} - \frac{M_A}{M_D} C_a - C_S \left(\frac{M_S}{M_D} + \frac{f}{f + \Psi} \right) \right] \left(\frac{M_N}{M_D} + \frac{\Psi}{f + \Psi} \right)^{-1}. \quad (2.33)$$

The atmospheric mass M_A is much less than the oceanic mass M_D and all the other terms apart from C_S do not depend on carbon concentrations, so the expression for C_N is of the form

$$C_N = -\phi C_S + c, \quad (2.34)$$

where c is a constant and ϕ is a function of f and Ψ , such that:

$$\phi = \left(\frac{M_S}{M_D} + \frac{f}{f + \Psi} \right) \left(\frac{M_N}{M_D} + \frac{\Psi}{f + \Psi} \right)^{-1}. \quad (2.35)$$

ϕ is only a function of positive quantities, so ϕ is *always positive*. From the relationship 2.34, C_N is proportional to C_S , coupled via the dynamical quantities f and Ψ . Therefore, if a forcing f is applied over the Southern box and results in a local change in C_S , ΔC_S ,

there will be a non-local change in the carbon concentration C_N , ΔC_N , in the Northern box which is proportional to ΔC_S such that:

$$\Delta C_N = - \left(\phi \Delta C_S + C_S \frac{\Delta \phi}{\Delta C_S} \Delta C_S \right) = - \left(\phi \Delta C_S + C_S \frac{\Delta C_S}{\Delta C_S / \Delta \phi} \right). \quad (2.36)$$

The proportionality between ΔC_N and ΔC_S is separated into two components:

- Since ϕ is always positive, $-\phi \Delta C_S$, the “re-distribution” term, reflects the fact if the carbon content in one part of the ocean changes, there must be an opposite change in carbon content in the rest of the ocean simply due to the conservation of carbon in the system, without any changes in dynamics, sources and sinks. The factor ϕ then simply measures the extent of the change in C_N .
- The term $-C_S \frac{\Delta C_S}{\Delta C_S / \Delta \phi}$ reflects that a forcing f can alter the large scale ocean circulation which in turn can affect the carbon content through the factor $\Delta C_S / \Delta \phi$. The factor $\Delta C_S / \Delta \phi$ reflects both dynamical and biogeochemical changes not addressed by the simple box model. The term can be either positive or negative, depending on the sign of $\Delta C_S / \Delta \phi$ in response to a particular forcing f . For example, ϕ decreases as f / Ψ increases according to eqn. 2.35. Therefore if the forcing f increases vertical mixing f more than the overturning Ψ , then carbon will be removed from the deep ocean and transferred to the surface ocean, thereby increasing C_S and giving rise to a negative $\Delta C_S / \Delta \phi$.

To relate the changes in carbon concentration ΔC_S and ΔC_N to changes in mean ocean $p\text{CO}_2$, $\overline{p\text{CO}_{2H}^o}$ and $\overline{p\text{CO}_{2E}^o}$ respectively, we use the definition of the Buffer Factor R (eqn. 1.6) which only holds for small perturbations in $p\text{CO}_2$:¹¹

$$\Delta p\text{CO}_2^o = R \frac{p\text{CO}_2^o}{C} \Delta C. \quad (2.37)$$

We take the spatial average of eqn. 2.37 separately in the region H and E to give:

$$\Delta \overline{p\text{CO}_{2H}^o} = R_H \frac{\overline{p\text{CO}_{2H}^o}}{C_H} \Delta \overline{C_H}, \quad (2.38)$$

$$\Delta \overline{p\text{CO}_{2E}^o} = R_E \frac{\overline{p\text{CO}_{2E}^o}}{C_E} \Delta \overline{C_E}, \quad (2.39)$$

where we have split the area integral on the right hand side to separate $\Delta \overline{C_H}$ and $\Delta \overline{C_E}$ in order to match these with C_S and C_N such that $\Delta \overline{C_H} \equiv \Delta C_S$ and $\Delta \overline{C_E} \equiv \Delta C_N$. Due to the small spatial variation of terms in eqns. 2.38 and 2.39, the error associated with this

¹¹we will later revisit over what range the assumptions are valid

operation is less than 1% in all MITgcm experiments used in this thesis. We can therefore write eqn. 2.36 as:

$$\Delta \overline{pCO_{2E}^o} = - \left(\phi + \overline{C_H} \frac{\Delta \phi}{\Delta \overline{C_H}} \right) \overline{R_E \frac{pCO_{2E}^o}{C_E}} \left(\overline{R_H \frac{pCO_{2H}^o}{C_H}} \right)^{-1} \Delta \overline{pCO_{2H}^o}, \quad (2.40)$$

$$\Delta \overline{pCO_{2E}^o} = \epsilon_f \Delta \overline{pCO_{2H}^o}. \quad (2.41)$$

The proportionality between $\Delta \overline{pCO_{2E}^o}$ and $\Delta \overline{pCO_{2H}^o}$ is expressed in terms of the parameter ϵ_f , which is a function of dynamical and biogeochemical quantities f , Ψ , R , pCO_2^o and C . The parameter ϵ_f is referred to as the non-local feedback parameter since it measures the influence of the local forcing f on the pCO_2^o in rest of the ocean basin via non-local dynamical and biogeochemical induced changes. Assuming $\Delta \overline{pCO_{2H}^o}$ and $\Delta \overline{C_H}$ are positive, the sign of f is determined entirely by $\Delta \overline{C_H}/\Delta \phi$ such that:

- $\epsilon_f > 0$: positive non-local feedback, $\Delta \overline{pCO_{2E}^o}$, so a positive ocean pCO_2 anomaly is created elsewhere in the basin due to a forcing f . ϵ_f can only be positive if $\Delta \overline{C_H}/\Delta \phi$ is negative and if its contribution is larger than the effect of the re-distribution term.
- $\epsilon_f \approx 0$: no non-local feedback, there is no change in ocean pCO_2 outside the forcing region. This can only occur if $\Delta \overline{C_H}/\Delta \phi$ is negative and if its contribution is equal to the effect of the re-distribution term or if there is no overturning circulation or mixing in the ocean at all.
- $\epsilon_f < 0$: negative non-local feedback, $\Delta \overline{pCO_{2E}^o}$ is the same sign as $\Delta \overline{pCO_{2H}^o}$ so a negative pCO_2^o anomaly is created elsewhere in the basin due to a forcing f . This can only occur if either $\Delta \overline{C_H}/\Delta \phi$ is positive or if the non-local feedback is dominated by the re-distribution term.

We can use the expression $\Delta \overline{pCO_{2E}^o} = \epsilon_f \Delta \overline{pCO_{2H}^o}$ to estimate the change in atmospheric pCO_2 due to the non-local feedback by substituting $\Delta \overline{pCO_{2E}^o}$ into eqn. 2.30 :

$$\Delta pCO_2^a = (1 + \epsilon_f \frac{P_E}{P_H}) P_H \Delta \overline{pCO_{2H}^o} + \Delta P_H \overline{pCO_{2H}^o} + \Delta P_E \overline{pCO_{2E}^o}. \quad (2.42)$$

The feedback induced by a forcing f on pCO_2^a is therefore split into 3 contributions:

- $P_H \Delta \overline{pCO_{2H}^o}$ measures the atmospheric pCO_2 feedback due to changes in **local** ocean pCO_2 due to a forcing f (thermodynamical, dynamical or biogeochemical),
- $\epsilon_f \frac{P_E}{P_H} P_H \Delta \overline{pCO_{2H}^o}$ measures the atmospheric pCO_2 feedback due to changes in **non-local** ocean pCO_2 due to a forcing f (through f/Ψ or biogeochemical),

- $\Delta P_H \overline{pCO_{2H}^o} + \Delta P_E \overline{pCO_{2E}^o}$ measures the atmospheric pCO_2 feedback due to changes in the local turbulent flux of carbon across the air-sea boundary without any changes in ocean carbon. This term only applies for a wind stress perturbation since ΔP_H and ΔP_E are zero for buoyancy and temperature changes.

Equation 2.42 will be applied to different regions throughout the remainder of this work to calculate specific local and non-local climate feedbacks.

Climate feedback of local Southern Ocean winds on atmospheric $p\text{CO}_2$

Westerly winds over the Southern Ocean and their impact on the atmosphere-ocean heat and carbon exchange have recently gained great scientific interest [Lauderdale et al., 2013, Lenton et al., 2013]. Southern Hemisphere westerly wind strength has been observed to increase over the last few decades [Swart et al., 2014] and is projected to rise further over the next century as the climate warms [Bracegirdle et al., 2013], potentially resulting in enhanced Ekman pumping and ventilation of the deep ocean. The strengthened ventilation leads to increased heat and carbon exchange between the ocean and the atmosphere [Sarmiento et al., 1998], further increasing global mean temperature.

In the contemporary climate, the Southern Ocean is the largest sink of atmospheric anthropogenic carbon [Khatiwala et al., 2009]. Using atmospheric CO_2 observations and atmospheric inversions between 1981 and 2004, Le Quere et al. [2007] find a reduction in the Southern Ocean carbon sink, attributed to the observed increase in westerly winds. However, other studies [Fay and McKinley, 2013, Majkut et al., 2014] apply different methods and find no significant change. Additional studies, such as Lovenduski and Ito [2009], analyse the evolving response of the Southern Ocean carbon sink to a changing climate and predict a reduction of the carbon sink in the coming century using a general circulation model. However, the lack of continuous observations and the transient response of the carbon cycle to simultaneously varying forcings (such as wind and buoyancy) limits the accuracy of the observed and projected estimates of the Southern Ocean carbon sink. In order to gain insight in the physical behaviour of the Southern Ocean carbon sink, we aim to derive a relationship between equilibrium Southern Ocean carbon and wind stress in terms of the physical parameters which determine the sensitivity of the response. Such a relationship can be used to understand and calculate the role of wind stress within the climate on the centennial timescales.

In this chapter, we focus on the local dynamical changes in the Southern Ocean where the wind stress perturbations occur (i.e. region H in Fig. 2.17). The first section aims

to derive a relationship for the sensitivity of atmospheric $p\text{CO}_2$ on Southern Ocean wind stress and the second section aims to use the relationship to investigate the behaviour of the Southern Ocean carbon system in climate-change scenarios, namely elevated atmospheric $p\text{CO}_2$ and surface temperatures. We will address the climate feedback that occurs due to dynamical and biogeochemical adjustments elsewhere in the ocean in chapter 4 (i.e. region E in Fig. 2.17).

3.1 Feedback of the natural Southern Ocean carbon flux

The starting point is to understand the response of ocean dynamics and the distribution of carbon due to the Southern Ocean wind stress perturbations. To that end, we employ the MITgcm and analyse the response of the system to perturbations in the zonal wind stress acting over the Southern Ocean channel (between 60° - 40°S).

3.1.1 Model response to increased zonal wind stress

The response of the MITgcm model to wind stress perturbations is discussed as generally as possible since the patterns of the response are common between all the perturbation experiments, while examples will be drawn from the $2 \times \tau_{SO}$ experiment on which detailed analysis was performed.¹ There is a modelled increase in the Ekman transport at the Northern edge of the channel as $\overline{\tau_{SO}}$ increases, where the overline indicates the area-mean between $60^\circ - 40^\circ\text{S}$ across the Southern Channel, shown in Fig. 3.2a. The Ekman transport is calculated as the meridional Eulerian volume transport across the basin between 0-100m depth at 40°S . Since the residual circulation also increases linearly with wind stress, there is limited eddy compensation. It must be pointed out that the linear response is an artefact of the coarse resolution of the model as a result of unresolved parametrised eddies. Munday et al. [2013] use a similar model set-up at $1/6^\circ$ resolution (albeit with a simpler wind stress forcing) which exhibits strong eddy saturation in the ACC, though compensation in the residual circulation is unaddressed. The bolus streamfunction Ψ^* increases as the isopycnal gradient steepens, but a constant GM diffusion coefficient does not provide the same compensation seen in models with resolved eddies or a variable GM coefficient [Deacu and Myers, 2005]. The limited eddy saturation and compensation is a limitation of the model and will be addressed later on. The changes in the residual, Eulerian and bolus streamfunctions are shown in Fig. 3.1.

¹The $2 \times \tau_{SO}$ experiment was chosen because the perturbation is strong enough for the response patterns to be clear, while limiting any potential non-linearities.

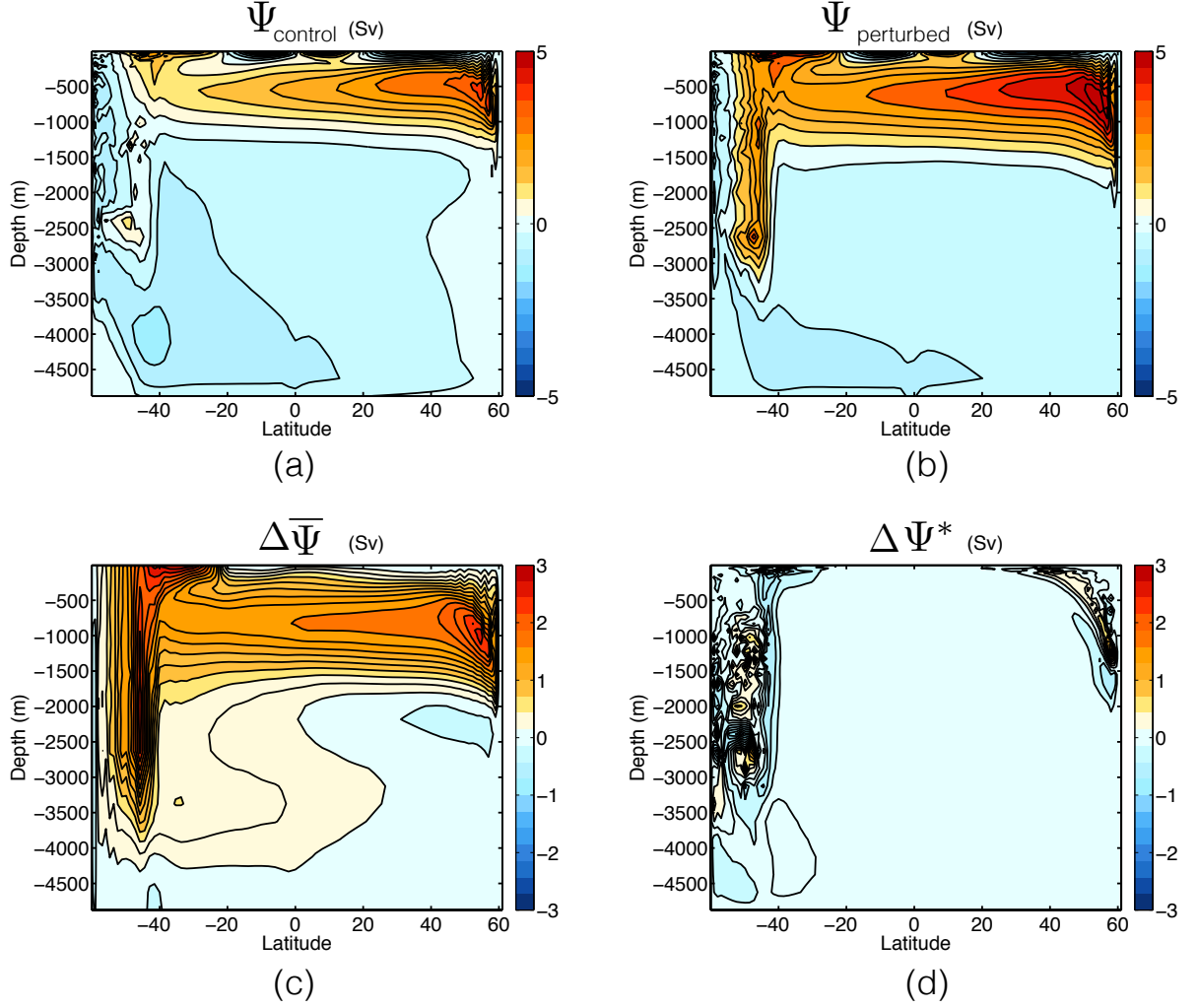


Figure 3.1: a) Control residual streamfunction, b) perturbed residual streamfunction, c) change in Eulerian and d) bolus streamfunctions after 400 years of a $2 \times \tau_{SO}$ perturbation.

The increase in channel Ekman transport feeds into a similar increase in the residual mean overturning. The westerly winds over the channel increase deep ocean upwelling as a result of the Ekman transport locally and the local increase in the residual circulation drives an increase in the MOC-like upper overturning circulation cell [Kuhlbrodt et al., 2007, Talley, 2013]. The effect is generally referred to as the “Drake Passage effect” [Toggweiler and Samuels, 1995, Gnanadesikan, 1999]. The strength of the MOC also increases roughly linearly within the range of wind stress perturbations studied, shown in Fig. 3.2b. Theories on simple scalings of the MOC strength, such as Gnanadesikan [1999] and Nikurashin and Vallis [2012], predict the depth of the MOC to scale with the MOC strength, the Southern Ocean wind stress and pycnocline depth,² however no clear trend

²Similar scalings have been used in other studies and usually take the form of $\Psi = \Delta_y b h^2 / f$ where

in the MOC depth is seen in the model experiments which is likely due to the vertical resolution of 100m at the relevant depth.

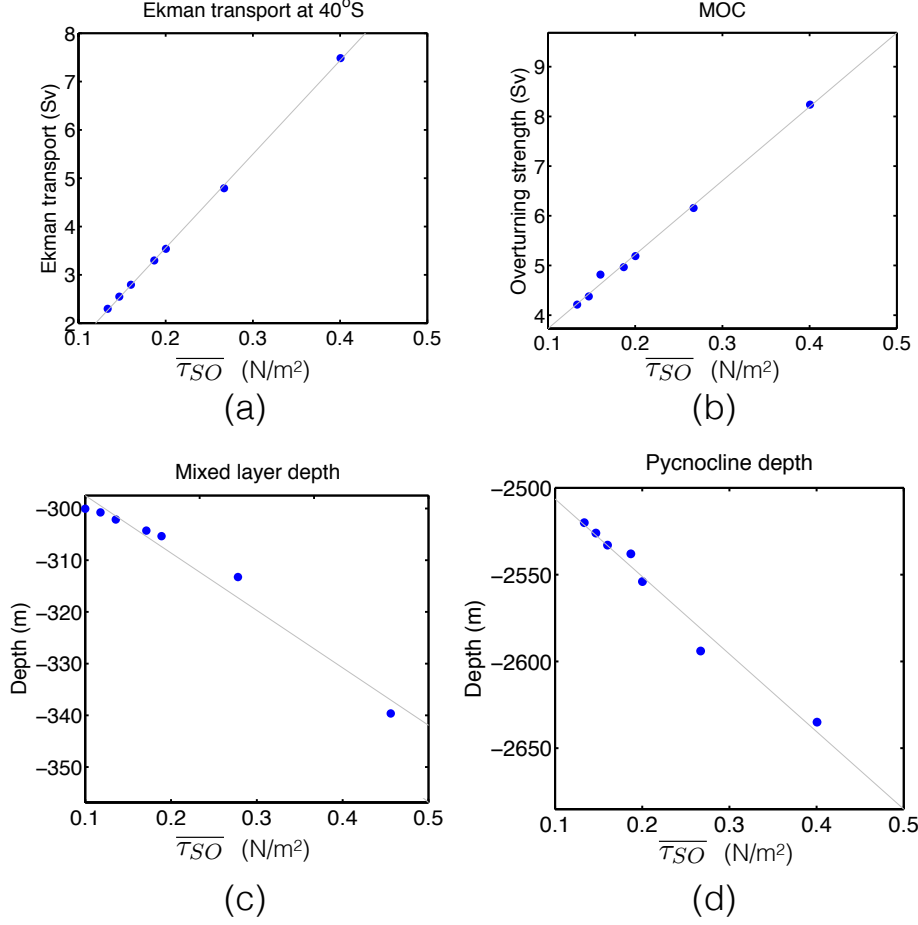


Figure 3.2: Dynamical quantities as a function of $\overline{\tau_{SO}}$ in MITgcm sector model: a) Strength of the Ekman transport at 40°S calculated as the meridional Eulerian volume transport between 0-100m depth across the whole width of the basin; b) Maximum of the overturning stream function between 350-1550m depth and 40°-60°N; c) Average mixed layer depth within the channel; d) Pycnocline depth at 40°S, calculated using the definition by Gnanadesikan [1999]. All quantities are taken as averages over the last 25 years of 400 year-long perturbation runs. Grey lines show a least-squares linear fit.

The increased wind-induced upwelling increases the surface density in the channel while the stronger Ekman transport causes increased production of Sub-Antarctic Mode Water, which manifests itself on the density structure in two crucial ways:

1. Steepening of the isopycnals within the channel (60°-40°S and 0-2500m depth) as

$\Delta_y b$ is the buoyancy window shared with the Southern Ocean, f is the Coriolis parameter and h is the pycnocline depth and also the depth scale of the MOC. The exponent of h is also uncertain because no agreed theory for the MOC exists.

dense water is pulled upwards;

2. Deepening of the isopycnals northwards in the interior of the basin, as the isopycnal depth in the mid-latitudes and Tropics is set by the isopycnal depth at the Northern edge of the channel and more deep water is lightened in the Southern Ocean due to the increased Ekman transport.

The change in the isopycnal depth and isopycnal slope in the Southern Ocean is illustrated by Fig. 3.3, which shows the change in the density structure induced by Southern Ocean winds.

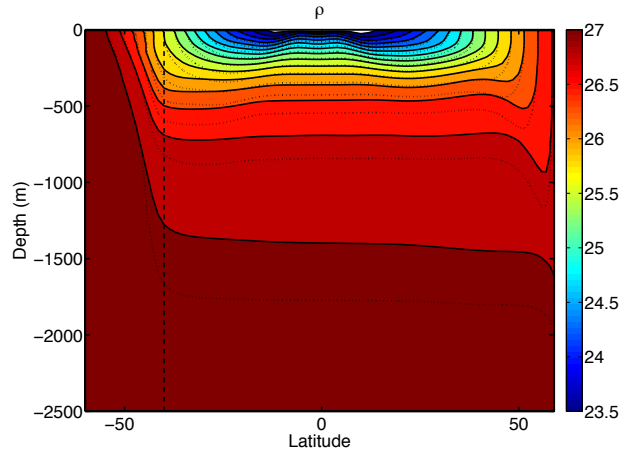


Figure 3.3: Zonal mean density (kg/m^3) in the MITgcm model. Solid lines show the control mean isopycnals and the dotted lines show the perturbed isopycnals after 400 years of a $2 \times \tau_{SO}$ perturbation.

Within the channel, the steepening of the density structure reduces the stratification (the vertical density gradient) resulting in an increase in mixed layer depth (Fig. 3.2c). Deepening of the mixed layer causes an increase in surface carbon concentration and a reduction in surface temperatures by mixing colder, more carbon rich waters with surface waters. The shifting isopycnal depth in the interior can be quantified by considering the pycnocline depth [Gnanadesikan, 1999] which marks the ventilation depth of carbon-rich waters such that:

$$z_\rho = \frac{\int_{-Z_D}^0 z [\sigma - \sigma(-Z_D)] dz}{\int_{-Z_D}^0 [\sigma - \sigma(-Z_D)] dz}, \quad (3.1)$$

where z_ρ is the pycnocline depth, σ is the potential density and Z_D is the depth of the basin (in this case, $Z_D = 5000\text{m}$). The values of σ and z_ρ are taken at the Northern edge

of the channel.³ Figure 3.2d shows z_ρ varying with $\overline{\tau_{SO}}$ in a linear fashion. However, the presence of eddies would result in a reduced sensitivity of the pycnocline to Southern wind stress [Munday et al., 2013].

The potential temperature response of the model reflects the changes in circulation and density structure. Figure 3.4b shows the mixed layer temperature change after 400 years of the $2 \times \tau_{SO}$ perturbation experiment. The increased upwelling in the channel leads to local cooling at the surface of up to -1°C . The zonal asymmetry North of the channel is the result of the increase in cold northwards moving waters, and a stronger western boundary current bringing warm tropical waters South. The zonally-averaged features, Fig. 3.4a, are also mainly a result of the circulation changes and have been extensively described in other studies [e.g. Lauderdale, 2010]. For example, the sub-surface maximum at around 25°S is the result of increased poleward ageostrophic flow caused by friction on a stronger ACC. The warmer temperatures further North of 25°S are a result of the increased MOC which circulates more water of a Northern origin instead of colder Antarctic Bottom Water (AABW), which becomes more isolated. Overall, the volume-integrated potential temperature in the basin increases with the wind stress strength.

Figure 3.4 shows a large difference in the DIC and temperature anomalies, because temperature and carbon have opposite vertical gradients (carbon concentration generally increases with depth and temperature decreases) and because carbon is affected by the presence of biology. The mixed layer carbon response (Fig. 3.4d) shows the increase in surface carbon in the Southern hemisphere associated with increased upwelling along with the zonal asymmetry associated with the increased sub-tropical gyre circulation. Zonally averaged anomalies (Fig. 3.4c) show an increase in carbon in the upper half of the MOC cell and a decrease in the lower half. The pattern is partly due to the isolation of carbon-rich AABW and partly because the overturning circulation stimulates global-scale mixing and therefore acts to erode the DIC depth gradient [Schmittner et al., 2007]. Carbon is therefore removed from the region between -1000 and -3000m depth and stored at shallower depths.

³Exactly which latitude(s) z_ρ is taken at is arbitrary. For this thesis, 40°S is chosen as it marks the imprint of the Southern channel density structure on the interior of the basin.

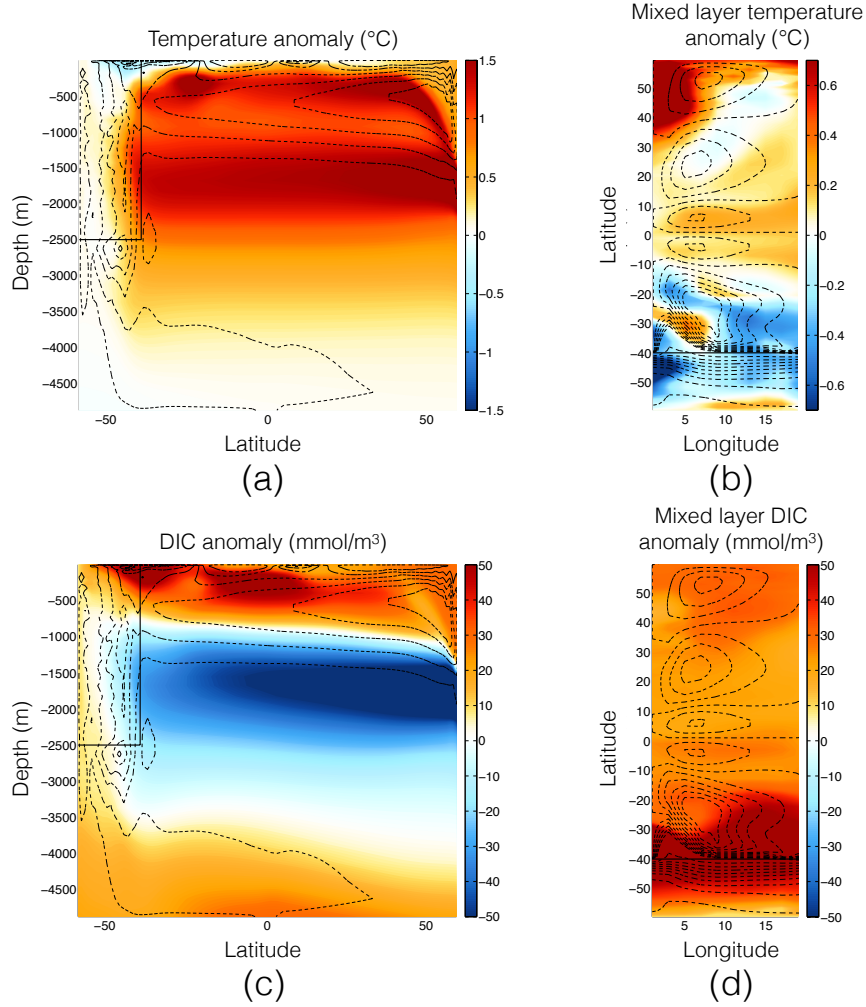


Figure 3.4: MITgcm output response to a $2 \times \tau_{SO}$ perturbation: Zonally averaged (a) potential temperature change and (c) dissolved inorganic carbon change as function of depth. Mean mixed layer (b) temperature change and (d) dissolved inorganic carbon change as function of longitude and latitude. All quantities are taken as the difference between an average over the last 10 years of a 400 year-long run and a control run average. In panels a and c, perturbed overturning streamfunction contours (dashed lines) are shown spaced 1 Sv apart. In panels b and d, contours of the perturbed barotropic streamfunction Ψ are shown. Solid black lines indicate the location of the Southern channel.

To understand the details of the pattern more clearly and for use in the next section, it is useful to split the change in preformed and regenerated carbon. The carbon can be further decomposed into saturated, soft, carbonate, anthropogenic saturated and disequilibrium carbon, but for the purposes of this study, the first order decomposition into preformed and regenerated carbon is sufficient (see Lauderdale [2010], Lauderdale et al. [2013], Munday et al. [2014] for a full description of the effects of Southern Ocean wind stress on each component.) The change in the preformed component shows a strong maximum at the surface in the channel (region 1 in Fig. 3.5a) as a result of the combined increased upwelling, reduced stratification and tilted isopycnals which have shifted northwards. The increase in preformed carbon below the surface is due to an increase in the disequilibrium component, which is to say that the increased circulation does not allow for surface outgassing of the excess upwelled carbon before subduction.

In contrast, the regenerated component shows a negative change in the channel up to 1500m-depth (region 2 in Fig. 3.5b). Increased upwelling generally brings more nutrients to the surface causing an increase in biological activity which would lead to an increase in regenerated carbon. As in the real ocean however, the biology at the surface of the Southern Ocean in the model is predominantly light limited, so that an increase in nutrient supply does not significantly affect biological productivity in the channel. Instead, the negative signal at mid-depth (1000-2000m) North of 40°S is advected upwards along isopycnals leading to the negative anomaly in the upper ocean. Between 1000-2000m depths, the increased MOC causes a reduction in regenerated carbon since the increased circulation causes water parcels to be less enriched by the “rain” of biological carbon from the surface: in a constant downward flux of biological carbon from above, or “rain”, the increased circulation reduces the residence time of water parcels in the “rain”, therefore causing a local reduction in biological carbon stores. This can be inferred from the tracer equation 2.24: for carbon $C = C_{pre} + C_{reg}$, the Lagrangian time derivative of preformed carbon is zero,⁴ i.e. $\frac{dC_{pre}}{dt} = 0$ (assuming adiabatic transport), since $F_{air-sea} = 0$. The only source of carbon at depth is F_{bio} , which only modifies regenerated carbon by definition. Assuming a horizontal velocity field $\mathbf{u}_h$ and neglecting diffusion, the rate of change

⁴Preformed carbon can be split into saturated and disequilibrium carbon. Saturated carbon depends on atmospheric CO₂, temperature and salinity, the latter two of which are assumed to be conserved below the thermocline. Disequilibrium carbon is set at the time of subduction.

of regenerated carbon is therefore given by:

$$\frac{\partial C_{reg}}{\partial t} = -\mathbf{u}_h \cdot \nabla C_{reg} + F_{bio}. \quad (3.2)$$

It follows that if the biological flux and horizontal gradient of C_{reg} remain constant, a local increase in the horizontal velocity $\mathbf{u}_h$ will reduce C_{reg} locally (as long as the term $\mathbf{u}_h \cdot \nabla C_{reg}$ is overall positive, which it is in this case). In the wind stress experiments, there is an increase in biological productivity in the mid-latitudes and the Tropics, but at depths between -1000 m and -3000 m the increase in horizontal advection, $\mathbf{u}_h \cdot \nabla C_{reg}$, is larger than the increase in F_{bio} causing a reduction in regenerated carbon concentration. In the equatorial region, between the thermocline and -1000m depth, the increase in F_{bio} is larger than the increase in $\mathbf{u}_h \cdot \nabla C_{reg}$, resulting in a build up of regenerated carbon. Isolation of the AABW cell also causes a build up of regenerated carbon (region 3 in Fig. 3.5b). It is worth noting that by definition, the regenerated carbon concentration is zero in the mixed layer. This definition means that when regenerated carbon is entrained through the bottom of the mixed layer, it is re-labeled as disequilibrium carbon [Ito and Follows, 2013], as in the Northern sub-polar region in region 4, Fig. 3.5b: the preformed anomaly originates from the upwelling of the equatorial regenerated carbon in the Northern sub-polar gyre.

The response of our model to wind stress is consistent with other studies [e.g., Jones et al., 2011, Lauderdale et al., 2013, Munday et al., 2013, 2014]. Comparing the MITgcm output to a more fully coupled global climate model such as HadGEM2-ES is difficult because of the multitude of atmospheric processes modified by the atmospheric forcing CO_2 , as is the case with observations. Nonetheless, studies such as Lovenduski et al. [2008] use hindcast simulations from a biogeochemical-physical model forced with NCEP atmospheric re-analysis and satellite data products between 1958 and 2004. Lovenduski et al. [2008] observes a similar pattern in natural carbon to our MITgcm simulations in the Southern Ocean, where the wind forcing is thought to be dominant: increased natural carbon at the surface North of the Southern Ocean and lower carbon concentrations between 200m and 2000m depth. To that end, we have confidence that the MITgcm model response is physical (as far as our current tools allow us to say).

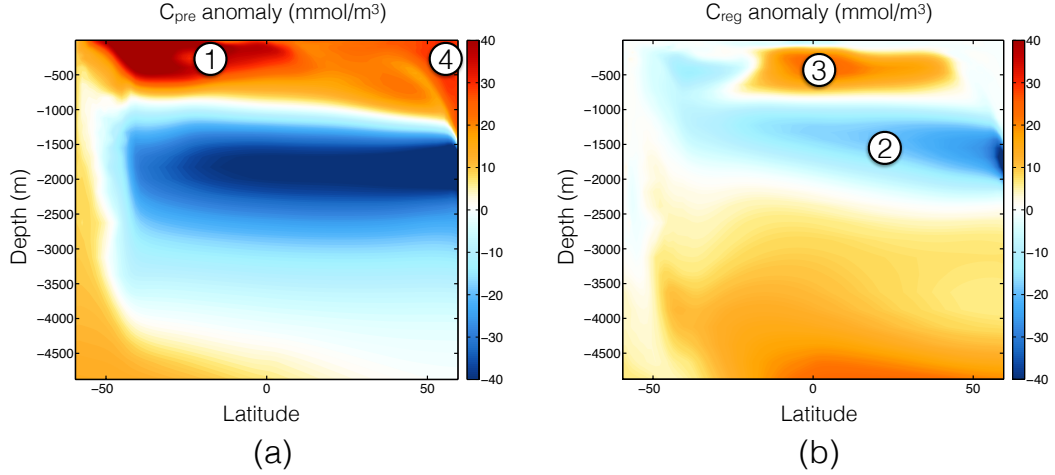


Figure 3.5: Zonal mean MITgcm output of the change in a) preformed and b) regenerated carbon in response to a $2 \times \tau_{SO}$ perturbation with freely evolving atmospheric CO_2 . All quantities are taken as the difference between an average over the last 10 years of a perturbed 400 year-long run and a control mean.

Figure 3.6 shows the anomaly in the Southern Ocean $p\text{CO}_2^o$ relative to the control mean, $\Delta \overline{p\text{CO}_2^o}$, for the set of wind stress experiments performed. While the experiment was run for 400 years, only the first 100 years are shown because the $\Delta \overline{p\text{CO}_2^o}$ is almost fully equilibrated at this time. The timescale of adjustment is the same for all experiments, with an e-folding time of 7-9 years and is predominately set by the vertical advection. Compared to the timescales listed in table 2.16b, the vertical and horizontal timescales are on the same order of magnitude as the adjustment timescale of $\Delta \overline{p\text{CO}_2^o}$ to wind

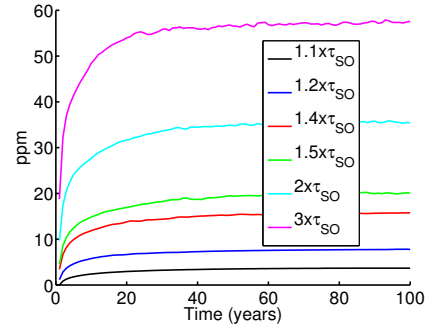


Figure 3.6: Southern Ocean change in $p\text{CO}_2^o$ averaged between $60^\circ - 40^\circ\text{S}$, $\Delta \overline{p\text{CO}_2^o}$, for range of τ_{SO} perturbations.

stress. Furthermore, the wind stress perturbation raises the surfaces of constant DIC in the Southern Ocean over a similar timescale. After the first 10 years, the increase in DIC due to advection causes a strong enough change in the Southern Ocean $p\text{CO}_2^o$ so that the air-sea flux anomaly begins to balance the advective DIC anomaly, which gives the profile of $p\text{CO}_2^o$ seen in the perturbation experiments. With an understanding of the physical response of the model, we can begin to express the sensitivity of the Southern Ocean $p\text{CO}_2^o$ to wind stress in terms of the relevant physical parameters and their effect on the atmospheric $p\text{CO}_2$ anomaly.

3.1.2 A theory for predicting the equilibrium climate feedback of Southern Ocean wind stress

To quantify the response of Southern Ocean mixed layer DIC and $p\text{CO}_2^o$, we start by using eqn. 2.26 which describes the integrated mixed layer carbon concentration C_h :

$$\frac{\partial C_h}{\partial t} = -\frac{C_h}{h}(w_h + u\frac{\partial h}{\partial x} + v\frac{\partial h}{\partial y}) + \frac{F_{air-sea}}{h} + \frac{[F_{bio}]_h}{h}. \quad (3.3)$$

To assess the impact of wind stress on C_h and $p\text{CO}_2^o$, we can express v , w_h and $F_{air-sea}$ in terms of the Southern Ocean wind stress, τ_{SO} . The residual horizontal velocity components (u, v) consist of a Eulerian and an eddy-induced contribution. The meridional Eulerian velocity can be approximated using the Ekman velocity $v_{ek} = \frac{\tau_{SO}}{f\rho_0 h}$, where f is the Coriolis parameter and ρ_0 is reference density [Gnanadesikan, 1999, Nikurashin and Vallis, 2012]. The eddy velocity v^* partially compensates the Ekman contribution and is assumed to be proportional to the slope of the isopycnals within the channel [Gent and McWilliams, 1990]. The slope of the isopycnals is also proportional to the Ekman transport so that it varies linearly with wind stress in our simulations. We therefore approximate v as $v = v_{ek} + v^* = (1 + e_{GM})\frac{\tau_{SO}}{f\rho_0 h}$ where e_{GM} is the proportionality constant between Ekman and eddy velocities (in the MITgcm set-up and in the majority CMIP5 models considered in this thesis, the strength of the Gent-McWilliams eddy diffusivity κ_{GM} is constant with wind stress as well so it is appropriate to use such an approximation [Meijers, 2014]). To determine w_h , we use the entrainment velocity, which measures the entrainment of water at the bottom of the mixed layer [Qiu and Kelly, 1993]. The entrainment velocity w_h is calculated through turbulent kinetic energy balance (for a full derivation, see appendix, section 7.4):

$$w_h = -\frac{2}{\alpha g h \Delta_z T} \left(m_0 u_*^3 - \alpha g h \frac{Q}{2\rho_0 C_p} \right), \quad (3.4)$$

where the parameters are:

- α : thermal expansion coefficient of water ($2.1 \times 10^{-4} \text{ K}^{-1}$),
- g : gravitational acceleration (9.8 m/s^2),
- $\Delta_z T$: temperature difference between the mixed layer and underlying water ($^{\circ}\text{C}$).

Because the concentration of DIC in the Southern Ocean mixed layer is determined by the DIC concentration of waters upwelled along isopycnals, we choose to estimate the temperature of “underlying water” at the mean depth of the isopycnals outcropping in the Southern channel, taken at the Northern edge of the channel (40°S). In our MITgcm set-up, this mean isopycnal depth at 40°S is roughly 1000m.

- m_0 : constant coefficient, $m_0 = 0.5$ following Davis et al. [1981] and Qiu and Kelly [1993], ⁵
- u_* : frictional wind velocity $u_* = \left| \frac{\tau_{SO}}{\rho_0} \right|^{1/2}$ (m/s),
- Q : net surface heat flux (W/m²),
- ρ_0 : reference density (1026 kg/m³),
- C_p : specific heat capacity of sea water at constant pressure (4000 J/kg/K).

The expression for w_h depends on the wind stress such that $w_h \propto \tau_{SO}^{3/2}$ and the sensitivity of w_h to wind stress depends on the variables Q , h and $\Delta_z T$, where $\Delta_z T$ is a measure of stratification. Lastly, the air-sea carbon flux in the Southern Ocean is given by $F_{air-sea} = K\tau_{SO}(pCO_2^a - pCO_2^o)$, where K is a constant and pCO_2^a and pCO_2^o are the atmospheric and oceanic pCO_2 , respectively (see eqn. 2.20). At equilibrium, $\frac{\partial C_h}{\partial t} = 0$, so we can solve equation 3.3 for pCO_2^o :

$$pCO_2^o = - \frac{C_h}{K} \left(\frac{2m_0}{\alpha g h \cdot \Delta_z T \rho_0^{3/2}} \tau_{SO}^{1/2} - \frac{Q}{\tau_{SO} \rho_0 C_p} + \frac{(1 + e_{GM})}{\rho_0 f h} \frac{\partial h}{\partial y} + \frac{u}{\tau_{SO}} \frac{\partial h}{\partial x} \right) + \frac{[F_{bio}]_h}{K\tau_{SO}} + pCO_2^a. \quad (3.5)$$

The first two terms on the right hand side arise from the expression for the vertical velocity w_h so they are both related to vertical mixing. However, we make the distinction between the mechanical mixing term (the first term) and the buoyancy-induced mixing term (second term). We are predominantly interested in the wind-induced mechanical mixing, since wind stress will affect the mechanical mixing more, we label the first and second terms the mixing and the radiative flux terms, respectively. The third term measures the influence of meridional advection, the fourth the influence of zonal advection and the fifth the influence of mixed layer biology on pCO_2^o . The last term represents equilibration with atmospheric pCO_2 .

We are interested in the area-average change in pCO_2^o which can be used to calculate a change in atmospheric pCO_2 induced by a wind stress anomaly. The area of outgassing of carbon in the MITgcm model is restricted to the area of positive vertical advection that is South of 40°S (in large-scale ocean models and the real ocean, outgassing generally occurs

⁵Davis et al. [1981] notes that m_0 is likely to depend on “sea state and surface roughness” and while the exact dependence is unknown, it is assumed to be constant.

South of 45°S [Majkut et al., 2014], therefore broadly in agreement with our model set-up). We therefore take a weighted average of eqn. 3.5 across the Southern Channel, between 40° – 60°S and 0° – 20°E, indicated by an overbar.⁶ The channel is zonally periodic so the mean zonal advection $\frac{u}{\tau_{SO}} \frac{\partial h}{\partial x}$ is zero when applying the averaging procedure. The averaged equation is given as:

$$\begin{aligned} \overline{pCO_2^o} = & - \overline{\left(\frac{C_h}{K}\right) \frac{2m_0}{\alpha_{gh} \Delta_z T \rho_0^{3/2}} \tau_{SO}^{1/2}} + \overline{\left(\frac{C_h}{K}\right) \frac{Q}{\Delta_z T \rho_0 C_p \tau_{SO}}} \\ & - \overline{\left(\frac{C_h}{K}\right) \frac{(1 + e_{GM})}{\rho_0 f h} \frac{\partial h}{\partial y}} + \overline{\frac{[F_{bio}]_h}{K \tau_{SO}}} + pCO_2^a, \end{aligned} \quad (3.6)$$

where pCO_2^a has no overbar since it is well-mixed in the atmosphere. We have so far assumed that the quantities Q , h and $\Delta_z T$ are constant with wind stress, which they are not. However, the variations of Q , h and $\Delta_z T$ with τ_{SO} are mostly second order. We therefore believe that the dominant contribution of τ_{SO} is captured in eqn. 3.6. This point is demonstrated by using the simplified version of eqn. 3.6, eqn. 3.7, to estimate Southern Ocean pCO_2 , which results in good agreement with model data, as will be shown later on in Fig. 3.8).

The ultimate goal is to derive an equation which will give the sensitivity of mean ocean pCO_2 to mean wind stress, i.e. $\frac{\partial \overline{pCO_2^o}}{\partial \overline{\tau_{SO}}}$. We would therefore like to split the area integral of the different terms to isolate $\overline{\tau_{SO}}$: e.g. $\overline{\left(\frac{C_h}{K}\right) \frac{2m_0}{\alpha_{gh} \Delta_z T \rho_0^{3/2}} \tau_{SO}^{1/2}}$ is split to give $\overline{\left(\frac{C_h}{K}\right) \frac{2m_0}{\alpha_{gh} \Delta_z T \rho_0^{3/2}}} \overline{\tau_{SO}}^{1/2}$. Figure 3.7 shows the magnitude of the terms in eqn. 3.6 as a function of area-mean wind stress $\overline{\tau_{SO}}$, both as they appear in eqn. 3.6 and with the area integral split to isolate $\overline{\tau_{SO}}$.

Figure 3.7 shows that splitting the area integrals for the spatial averages in eqn. 3.6 results in large errors for all terms that depend on τ_{SO} except the vertical mixing term. In the radiative flux and biology terms, spatial correlations with the wind stress cause large errors if the integral for τ_{SO} is taken separately. The error in the vertical mixing term is less than 5%. We therefore only preserve the split integral for the vertical mixing term.

Figure 3.7 shows that in the control state ($\overline{\tau_{SO}} = 0.13 \text{ N/m}^{-2}$), Southern Ocean pCO_2^o is predominantly set by vertical mixing, biology and meridional advection, while the radiative flux term is negligible. However, as wind stress increases, the change in equilibrium

⁶To take the average, we apply the integral $\int_{60S}^{40S} \int_{0E}^{20E} (\text{eqn. 3.5}) dx dy / \int_{60S}^{40S} \int_{0E}^{20E} dx dy$.

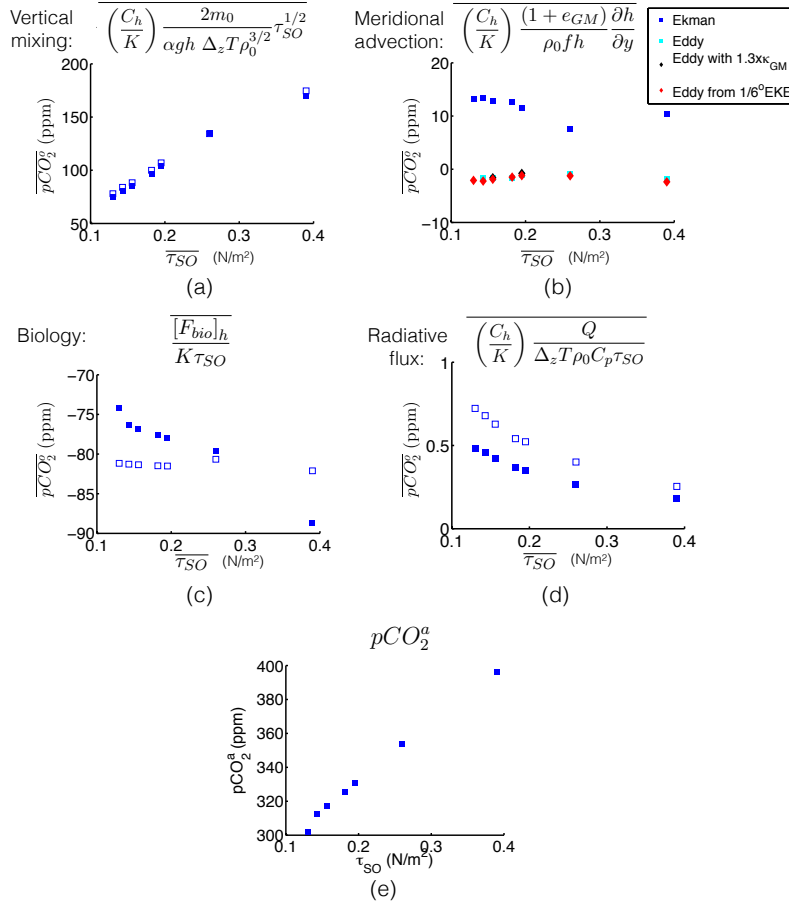


Figure 3.7: Contributions of a) mixing, b) meridional advection, c) biology and d) radiative flux terms to the estimate of Southern Ocean $p\text{CO}_2$ (eqn. 3.7) as a function of Southern Ocean wind stress. Empty squares show the terms as they appear in eqn. 3.6 and the filled squares show the terms with the average split to isolate $\overline{\tau_{SO}}$. Panel (b) shows the Ekman (dark blue) and eddy (cyan) contributions, as well as the eddy contributions in experiments with the GM coefficient increased by 30% (black diamonds) and with v^* estimated from eddy-permitting Southern Ocean wind stress experiments performed by Munday et al. [2013] (red diamonds). Panel (e) shows atmospheric $p\text{CO}_2$ as a function of wind stress.

$\overline{p\text{CO}_2}$ induced by the wind stress is driven by the vertical mixing. The meridional advection term is roughly constant with wind stress due to linear dependence of the Ekman velocity and air-sea flux on τ_{SO} . The τ_{SO} -terms cancel in eqn. 3.5 and the meridional advection term therefore does not explicitly depend on τ_{SO} .

While eddies in the Southern Ocean depend on wind stress (e.g. Munday et al. [2013]), eddies are unresolved in our MITgcm set-up. Instead, eddies are parametrised as an isopycnal diffusion term with the strength of diffusivity κ_{GM} as a constant. Other studies suggest that κ_{GM} should increase as the wind stress increases, to model the increasing eddy activity [Deacu and Myers, 2005]. To address this concern, simulations were run whereby

κ_{GM} was increased by 30% and the strength of the eddy-induced meridional advection term in eqn. 3.6 diagnosed. Additionally, Munday et al. [2013] performed eddy-permitting ($1/6^\circ$ horizontal resolution) Southern Ocean wind-stress perturbation experiments in a similar MITgcm configuration as the one used in this thesis. The basin-averaged Eddy Kinetic Energy (EKE) is found to be linearly dependent on the maximum Southern Ocean wind stress. We therefore use the diagnosed EKE as a function of Southern Ocean wind stress in the Munday et al. [2013] experiments to estimate v^* and the eddy-induced meridional advection term in eqn. 3.6. The result of these calculations is shown in Fig. 3.7b and we find that in all cases, the net contribution of eddies towards Southern Ocean $p\text{CO}_2$ is small, less than 5 ppm for wind stress values of up to $\overline{\tau_{SO}} = 0.4 \text{ N/m}^{-2}$.

The mixed layer biology $[F_{bio}]_h$ has no explicit dependence on wind stress, however, the biology term $\frac{[F_{bio}]_h}{K\tau_{SO}}$ in eqn. 3.5 varies as the inverse of τ_{SO} . Figure 3.7c shows that the biology term is roughly constant with $\overline{\tau_{SO}}$, which means that $\overline{[F_{bio}]_h}$ also increases linearly with wind stress. Overall, the biology term is therefore independent of $\overline{\tau_{SO}}$, although this may vary in different model set-ups. The net result is the scaling:

$$\overline{pCO_2^o(\tau_{SO})} = -C_h \frac{2m_0}{K\alpha gh \cdot \Delta_z T \rho_0^{3/2}} \overline{\tau_{SO}}^{1/2} + pCO_2^a + \mathfrak{F}(v, Q, F_{bio}), \quad (3.7)$$

where $\mathfrak{F}(v, Q, F_{bio})$ is a term capturing the effects of Ekman, eddies, the radiative flux and mixed layer biology, but which does not vary significantly with τ_{SO} :

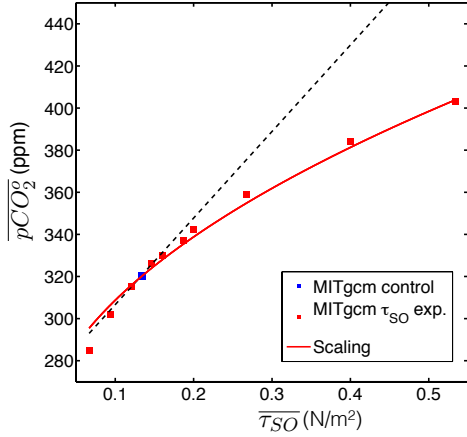
$$\mathfrak{F}(v, Q, F_{bio}) = \left(\frac{C_h}{K}\right) \frac{Q}{\Delta_z T \rho_0 C_p \tau_{SO}} - \left(\frac{C_h}{K}\right) \frac{(1 + e_{GM})}{\rho_0 f h} \frac{\partial h}{\partial y} + \frac{[F_{bio}]_h}{K\tau_{SO}}. \quad (3.8)$$

The mixed layer depth h can be output from the model and $\Delta_z T$ is calculated as described above. To determine the appropriate value for K , eqn. 3.7 is solved for K using control state values for all other parameters. The value of K is then calculated as $K\overline{\tau_{SO}} = 42 \text{ cm/hr}$, which is within the range suggested by previous studies [Wanninkhof, 1992, Wanninkhof et al., 2009, Rutgersson et al., 2011]. While K is specified in the model, diagnosing K as described allows for different methods for calculating the mixed layer depth h and $\Delta_z T$ to give consistent results. The values of h , $\Delta_z T$ and K are calculated at each grid point and then averaged over the Southern Ocean.

Equation 3.7 demonstrates that the equilibrium $\overline{pCO_2^o}$ in the Southern Ocean varies as the square root of the Southern Ocean wind stress. The key variables in determining the sensitivity of $\overline{pCO_2^o}$ to the wind stress are C_h , h and the stratification $\Delta_z T$, while the

contributions of the Ekman, eddy transport and biology do not contribute to the sensitivity of the equilibrium $\overline{pCO_2^o}$ to a perturbation in wind stress. A higher background DIC concentration C_h will result in upwelling of more carbon, causing a larger change in $\overline{pCO_2^o}$. Entrainment of carbon-rich water into a shallow mixed layer will increase concentrations more than a deep mixed layer due to the inertia of the layer. Furthermore, strong stratification $\Delta_z T$ inhibits the upwelling and mixing of deep waters into the surface layer so will dampen the response of $\overline{pCO_2^o}$ to the wind stress perturbation.

In practice, Fig. 3.8 shows the results from the MITgcm wind stress perturbation experiments (squares) along with the calculations with eqn. 3.7 (solid red line). Figure 3.8



(a)

Parameter	MITgcm control
$\overline{C_h}$	2170 mmol/m ³
$\overline{h}$	300 m
$\overline{\Delta_z T}$	4.7 °C
$\overline{\tau_{SO}}$	0.13 N/m ²

(b)

Figure 3.8: a) $\overline{pCO_2^o}$ (after 400 years) in the Southern Ocean as a function of $\overline{\tau_{SO}}$. The blue square shows the control value, the red squares show MITgcm perturbation experiment output and the solid red line shows the calculation using the derived scaling, eqn. 3.7. The dashed line shows a linear Taylor expansion around the control run value. b) Table showing the area mean values of the variables used in the scaling equation as diagnosed from the MITgcm control run.

shows that the predicted Southern Ocean pCO_2^o values using the scaling closely matches the MITgcm results which supports the validity of equation 3.7. To test the robustness of the scaling with respect to the geometry of the problem, we perform the same set of wind perturbation experiments, as described in section 2.3.4 with a) a rectangular basin 40°-wide in longitude and b) with a triangular wind stress profile between 60° and 40°S. In both instances, the response in $\overline{pCO_2^o}$ with $\overline{\tau_{SO}}$ is roughly similar in magnitude and

follows a square root relationship (for details, see appendix, sections 7.5 and 7.6). The result described by eqn. 3.7 therefore holds regardless of the choice of basin size and wind stress profiles. We have therefore shown that the Southern Ocean $\overline{pCO_2^o}$ varies as the square root of the mean wind stress and is driven by vertical mixing.

The influence of the stratification, carbon content and wind stress determines not only the Southern Ocean $\overline{pCO_2^o}$ but the net flux of carbon between the ocean and atmosphere in the Southern Ocean. We can therefore estimate the *local*⁷ climate-carbon feedback due to the wind-induced ocean pCO_2 change by using eqn. 2.42:

$$\Delta pCO_2^a = P_{SO} \Delta \overline{pCO_{2SO}^o} + \Delta P_{SO} \overline{pCO_{2SO}^o} + \Delta P_E \overline{pCO_{2E}^o}. \quad (3.9)$$

where Δ indicates the anomaly with respect to the control mean, the subscript *SO* refers to the Southern Ocean and the subscript *E* refers to the rest of the ocean basin outside of the Southern Ocean. P_{SO} and P_E are the area fractions of the Southern Ocean and the rest of the basin weighted by the local wind stress (see section 2.4.2). To use eqn. 3.9, we require the change in $\overline{pCO_2^o}$ due to a wind stress perturbation. Expressing eqn. 3.7 in terms of perturbations to the control state gives:

$$\Delta \overline{pCO_2^o} = -C_h \frac{2m_0}{K\alpha gh \cdot \Delta_z T \rho_0^{3/2}} \Delta(\overline{\tau_{SO}}^{1/2}) + \Delta pCO_2^a, \quad (3.10)$$

where we assume that the parameters C_h , h and $\Delta_z T$ do not change as a result of the wind stress perturbation⁸ and that $\Delta \mathfrak{F}(v, Q, F_{bio}) = 0$ since the term $\mathfrak{F}(v, Q, F_{bio})$ is independent of wind stress. Substituting for $\Delta \overline{pCO_2^o}$ in eqn. 3.9 and solving for ΔpCO_2^a then gives the final feedback equation:

$$\Delta pCO_2^a = \frac{P_{SO}}{1 - P_{SO}} C_h \frac{2m_0}{K\alpha gh \Delta_z T \rho_0^{3/2}} \Delta(\overline{\tau_{SO}}^{1/2}) + \frac{\Delta P_{SO}}{1 - P_{SO}} \overline{pCO_{2SO}^o} + \frac{\Delta P_E}{1 - P_{SO}} \overline{pCO_{2E}^o}, \quad (3.11)$$

To validate eqn. 3.11 we compare it with the MITgcm output from the wind stress perturbation experiments. However, we want to estimate the ΔpCO_2^a due to local Southern Ocean effects alone, so we remove the ΔpCO_2^a signal from non-local effects (the procedure for this calculation is outlined in the appendix, section 7.8). The equilibrium atmospheric ΔpCO_2^a due to Southern Ocean wind stress perturbations is shown in Fig. 3.9. For comparison, both the full atmospheric ΔpCO_2^a from the MITgcm experiments as well as the

⁷We only calculate the feedback due to local change in Southern Ocean pCO_2 , therefore assuming $\Delta \overline{pCO_{2E}^o} = 0$ for now.

⁸This assumption is not strictly true, but the scaling is designed to be used in climate models and the real ocean, where we know the values of C_h , h and $\Delta_z T$ at the time $\Delta \overline{pCO_2^o}$ is evaluated.

ΔpCO_2^a with non-local effects removed are shown. There are significant differences between the full and local ΔpCO_2^a . The source of the additional non-local ΔpCO_2^a is addressed in chapter 4.

Figure 3.9 also shows the calculation using only the $\Delta(\overline{\tau_{SO}}^{1/2})$ term in eqn. 3.11, which reduces the estimate of ΔpCO_2^a by 9%. The contributions of the ΔP_{SO} - and ΔP_E -terms are therefore small and the atmospheric carbon-climate feedback due to wind stress is predominantly set by the dynamical change in Southern Ocean pCO_2 due to vertical mixing.

The estimation of ΔpCO_2^a using eqn. 3.11 was done for the MITgcm control state, as well as for the CMIP5 models HadGEM2-ES and IPSL-CM5, and while the input parameters vary between all the models, the net results are within the same order of magnitude. However, the local and area-mean values of $\Delta_z T$ and h can vary as much as 200 – 300% between models, resulting in similar differences in the calculated ΔpCO_2^a . For example, a $\Delta\overline{\tau_{SO}}$ of 0.2 N/m² results in an atmospheric pCO_2 change of 16 ppm for IPSL-CM5 and 48 ppm for HadGEM2-ES, a 300% difference. The atmospheric pCO_2 feedback therefore varies significantly between models and depends on the physical parameters setting Southern Ocean stratification. The effective area fraction term P_{SO} was found to range between 0.27–0.37 for MITgcm for $1-3 \times \tau_{SO}$ scenarios. For the CMIP5 models, the effective area fraction ranges between 0.43 – 0.47 for the pre-industrial state, which shows the impact Southern Ocean pCO_2 has on atmospheric carbon concentrations, despite only covering 23% of the world’s ocean, due to the high local wind stress.

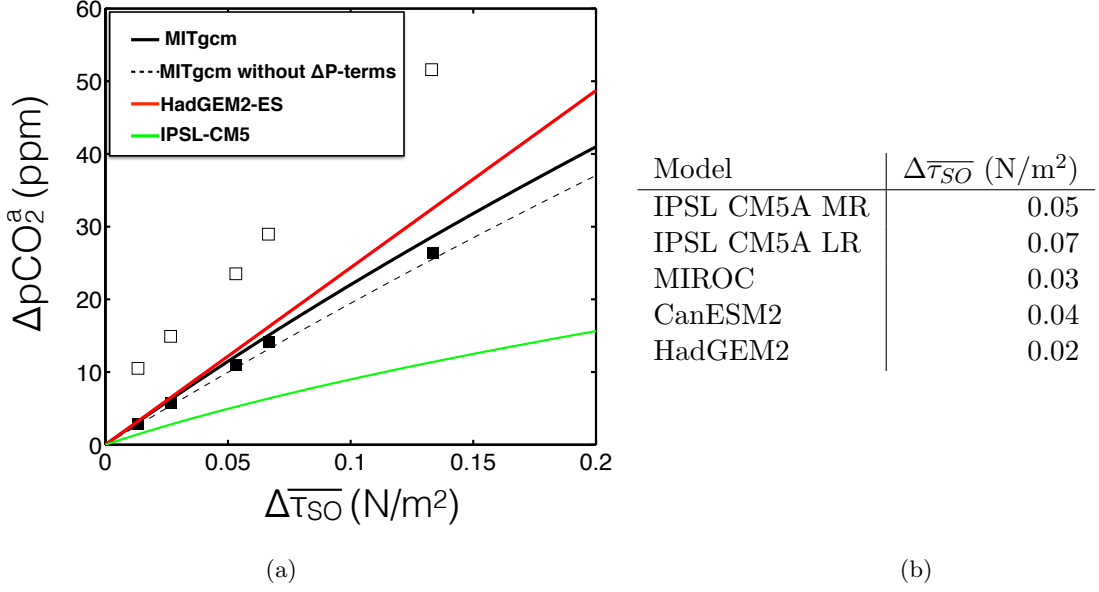


Figure 3.9: a) Predicted change in equilibrium concentration of atmospheric CO_2 due to a perturbation in mean Southern Ocean wind stress. The response is calculated using only changes in Southern Ocean pCO_2^a . The solid lines show the atmospheric feedback as predicted using control parameter values for the MITgcm (black), HadGEM2-ES (red) and IPSL-CM5 (green). The squares show data from the MITgcm wind stress perturbation experiments: empty squares show the full ΔpCO_2^a modelled, while the filled squares show the ΔpCO_2^a due to changes in Southern Ocean pCO_2^a only. The dashed line shows the predicted ΔpCO_2^a using only the $\Delta \overline{\tau}_{SO}^{1/2}$ -term in eqn. 3.11. b) Table of $\Delta \overline{\tau}_{SO}$ for CMIP5 $4 \times CO_2$ experiments (between 70 – 45°S)

We can illustrate the upper bound of expected wind stress changes by considering the CMIP5 $4 \times CO_2$ experiments, where the average $\Delta \overline{\tau}_{SO}$ is 0.042 N/m². The corresponding equilibrated atmospheric response is ΔpCO_2^a is 4 – 10 ppm. Swart and Fyfe [2012] show that CMIP5 models predict a 10% increase in Southern Hemisphere wind stress at the end the 21st century under an RCP8.5 scenario from current levels, which would give an equilibrium ΔpCO_2^a of ~ 5 ppm based on our results.

However, Swart and Fyfe [2012] also point out that CMIP5 models strongly underestimate historical trends in $\overline{\tau}_{SO}$: between 1979 and 2010, CMIP5 calculate a trend of 0.002 – 0.003 Pa, while reanalysis averages are closer to 0.01 Pa, though with large error bars. A ΔpCO_2^a of 5 ppm assuming equilibration at the end of the 21st century is therefore a lower estimate. Friedlingstein et al. [2006] use C4MIP models to estimate the global carbon climate feedback in 2100 to most likely be between 50 and 100ppm. Based on our approach, the ΔpCO_2^a due to Southern Ocean wind stress is therefore likely to be

5 – 10% of the total. This estimate only takes into account the response of natural carbon in the absence of other climate forcings such as surface warming. In the context of climate change, the Southern Ocean is one of the main sinks of anthropogenic carbon, so we are also interested in the response of uptake of anthropogenic carbon and how the sink may change as the climate warms. We can apply our scaling to help us understand the future magnitude of the Southern Ocean carbon sink in the context of climate change.

3.2 The wind stress feedback in future scenarios

3.2.1 The anthropogenic carbon sink

In the contemporary climate, the Southern Ocean is a net sink of carbon since current anthropogenic atmospheric CO_2 levels are high enough to drive a flux of anthropogenic carbon into the Southern Ocean. By altering the background Southern Ocean $p\text{CO}_2^o$, the anthropogenic sink could potentially be reduced. To investigate the effects of Southern Ocean wind stress on the anthropogenic carbon flux and to interpret our results in the context of the Southern Ocean carbon sink [Le Quere et al., 2007, Sabine et al., 2004], we run a new set of forced MITgcm experiments (see Fig. 3.10). In the runs, we increase $\overline{\tau_{SO}}$ as before at $t = t_0$, and simultaneously introduce an anthropogenic CO_2 pulse by increasing the initial control $p\text{CO}_2^a$ of 300ppm by 260ppm (to give 560ppm, equivalent to a CMIP5 $2\times\text{CO}_2$ scenario). Such an atmospheric pulse represents an instantaneous input of 27 Gton of carbon. The system is left to adjust in order to monitor the transient adjustment of the anthropogenic carbon as the ocean responds to the wind stress anomaly (which is held fixed for 400 years). The anthropogenic carbon pulse experiments allow us to calculate the anthropogenic air-sea flux of carbon, $F_{air-sea}^{ant}$, by subtracting the natural carbon flux (from the wind-stress perturbation experiments alone) from the total flux in the anthropogenic carbon runs.

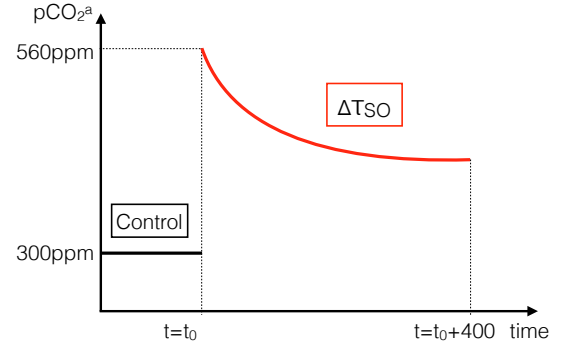


Figure 3.10: Schematic illustrating the evolution of atmospheric $p\text{CO}_2$ in wind-stress perturbation experiments with an anthropogenic pulse of 260ppm at time $t = t_0$. The atmospheric pulse is introduced at the time of the wind-stress perturbation, after which time the perturbation in wind-stress is held for 400 years during which time the atmospheric $p\text{CO}_2$ evolves freely.

Spatially, the differences are concentrated in the Northern and Southern high latitudes where the majority of ocean anthropogenic carbon uptake occurs [Khatiwala et al., 2009]. Figure 3.11a shows the temporal evolution of atmospheric carbon concentrations in the same range of $\overline{\tau_{SO}}$ perturbations studied previously. Because the Southern Ocean $p\text{CO}_2^o$ is elevated by the wind stress anomaly, the sink of carbon is reduced so that more remains in the atmosphere and the timescale of ocean carbon uptake increases.

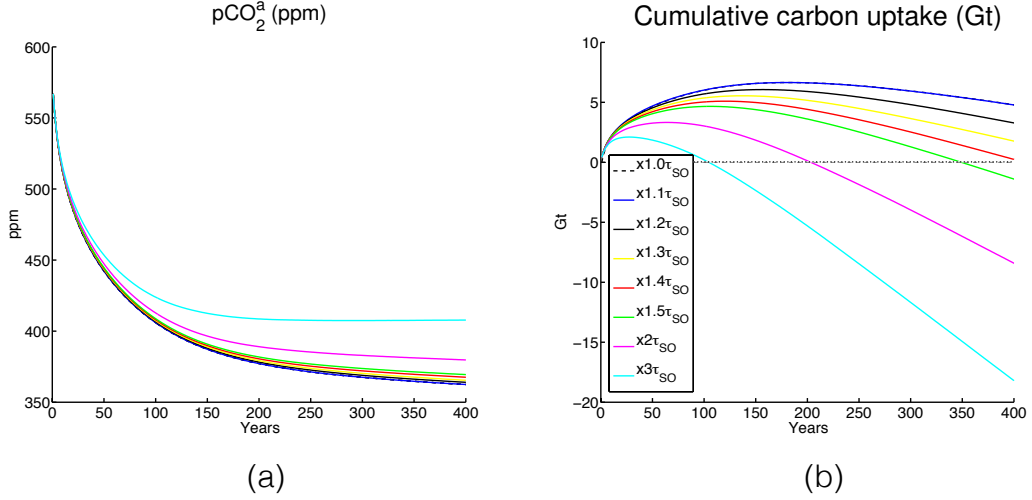


Figure 3.11: a) Time evolution of $p\text{CO}_2^a$ and b) cumulative Southern Ocean air-sea flux of carbon in $\Delta\tau_{SO}$ with anthropogenic pulse experiments performed with MITgcm. The dotted line in panel (b) shows the zero value.

While our study focusses on the equilibrium Southern Ocean flux, current observations are inherently transient. Therefore we can examine the transient response of the anthropogenic carbon sink in the MITgcm model and relate the changes in the sink to observations based studies. We can calculate that after the first 5 years of adjustment, the mean air-sea flux of anthropogenic carbon in the Southern Channel is reduced by a constant amount $0.04 \text{ mol/m}^2/\text{year}$ for every 0.012 N/m^2 increase in $\Delta\overline{\tau_{SO}}$ over the area $60^\circ - 20^\circ\text{N}$ (to accurately estimate the impact of Southern Ocean wind stress on the Southern Ocean carbon sink, we will consider the area from $60^\circ - 20^\circ\text{N}$ for this section, which includes the Southern channel as well the area of natural carbon uptake just North of the channel). After 10 years, a 0.012 N/m^2 perturbation in $\Delta\overline{\tau_{SO}}$ gives rise to a 3% decrease in the sink. Swart and Fyfe [2012] note that between 1979 and 2010, reanalysis suggests a trend in mean Southern Ocean wind stress of roughly 0.01 N/m^2 per decade, although the uncertainty is large. Over the 3 decades under investigation, a trend of 0.01 N/m^2 per decade suggests a 0.03 N/m^2 change in $\overline{\tau_{SO}}$ which could therefore reduce the transient

Southern Ocean carbon sink by at least 3% per decade. Sabine et al. [2004] estimate the Southern Ocean carbon sink reduction by using reanalysis and model simulations between the years 1981 and 2004. They calculate that while the expected trend in air-sea carbon flux due to the atmospheric trend is 0.051 Pg C/year, the ‘observed’ or modelled trend is close to 0.031 Pg C/year, suggesting a 40% change. However, later studies by Fay and McKinley [2013] find no significant difference in ocean and atmosphere trend at all and Majkut et al. [2014] even note an increase in the Southern Ocean sink due to surface cooling, so the observations still leave us unclear as to how the real Southern Ocean is adapting to the changing climate.

The cumulative effects of the reduced carbon sink become more apparent with time. Figure 3.11b shows the cumulative air-sea carbon flux over the area $60^\circ - 20^\circ\text{N}$. In the Southern Ocean, there is a flux of natural carbon from the ocean into the atmosphere, and an anthropogenic flux in the opposite direction. Initially, the anthropogenic carbon flux is larger than the natural flux such that the Southern Ocean acts as a sink. However, with time, the atmospheric anthropogenic carbon anomaly decays which reduces the flux of anthropogenic carbon into the ocean and the natural carbon flux becomes stronger, causing the Southern Ocean to be a net source of carbon. The stronger the wind stress perturbation, the sooner the turnaround point occurs. While the dynamical differences between wind stress scenarios are small, the differences in the timing of the switch in sign of the cumulative carbon flux are large. A $\Delta\overline{\tau_{SO}}$ of 0.012 N/m² over the control value gives a 50 year decrease in the duration for which the Southern Ocean remains a sink of carbon. Time delays for 50 years in the absorption of atmospheric could have important implications for the global temperature response, especially when considering emission mitigation targets that aims to limit peak global warming. Such delays in the switch of the sign of the carbon sink illustrate how Southern Ocean wind stress-induced uncertainties in the air-sea carbon flux affect the timing of peak global warming, as was introduced in the simple model study in section 2.1.

Moreover, since the carbon flux is dictated by the local $\overline{pCO_2}$, it follows that the cumulative flux will obey a square root relationship with respect to wind stress. Figure 3.12 shows that the Southern Ocean cumulative carbon flux also scales as the square root of the wind stress. The response is shown after 10 and 300 years so the results hold

regardless of the timescale, which is only true if we consider the local effect of the Southern Ocean alone.⁹

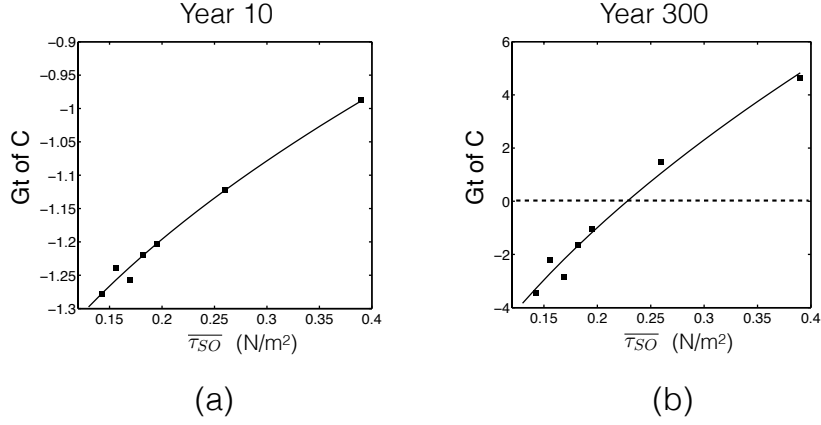


Figure 3.12: Cumulative carbon flux $60^\circ - 20^\circ N$ in MITgcm after a 260ppm anthropogenic carbon pulse as a function of wind stress, for a) 10 and b) 300 years after the perturbations. Squares show MITgcm output, while the solid lines show fits to a $\overline{\tau_{SO}}^{1/2}$ trend.

A useful feature of our wind stress scaling is that it identifies the key physical processes and parameters that determine the sensitivity of the $\overline{pCO_2^o}$ response to wind stress changes. We can therefore use the scaling to calculate the sensitivity of the carbon flux in different climate models that might have similar wind stress profiles, but different ocean states that will affect the strength of the feedback. Moreover, we have only used pre-industrial values for the parameters C_h , h and $\Delta_z T$ in equation 3.7 so far. As the climate warms however, the value of these parameters will change due to forcings other than wind stress, changing the flux and the uptake timescale of Southern Ocean carbon. To test the robustness of the scaling, we can therefore use the MITgcm to test the sensitivity of $pCO_2^o(\overline{\tau_{SO}})$ to C_h , h and $\Delta_z T$.

3.2.2 Sensitivity to Southern Ocean stratification

The parameters of the surface ocean which set the feedback of Southern Ocean wind stress respond to the changing climate. C_h is the most sensitive to atmospheric CO_2 perturbations, while h and $\Delta_z T$ are most sensitive to surface buoyancy forcing and circulation changes. A perturbation in atmospheric pCO_2 will cause the ocean pCO_2 to track the changes in the atmosphere, maintaining a constant disequilibrium given no (thermo)dynamical changes [Fay and McKinley, 2013]. However, changes to the large

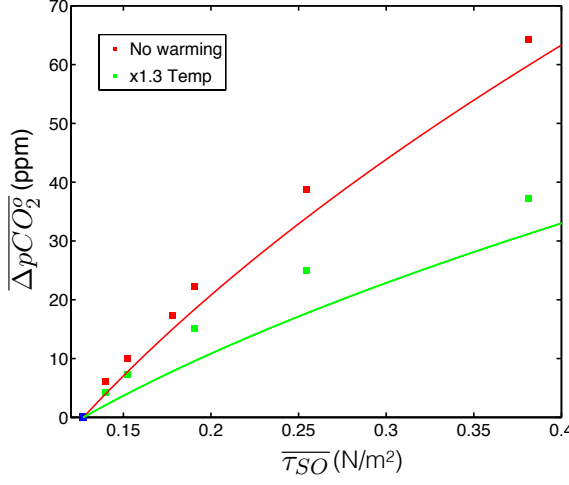
⁹We will see later that non-local effects elevate atmospheric pCO_2 further, which causes a deviation from the trend for timescales longer than 20 years. The details of how the non-local effects are removed and their impacts will be discussed later.

background DIC concentration are relatively small and will not significantly affect dynamical feedbacks (this is discussed more later in this section and in chapter 6). Under changes in buoyancy forcing however, the parameters h and $\Delta_z T$ will be affected. Changes in the climate such as surface warming also affect quantities such as the radiative flux Q and the horizontal transport of carbon, though the contribution of these terms towards ocean $p\text{CO}_2$ is small, as shown in section 3.1.2.

To test the dependence of the climate feedback on these parameters, further experiments were run whereby an increase in surface heating Q of 30% is applied across the whole domain for 400 years¹⁰ after which a perturbation in $\overline{\tau_{SO}}$ from 10% to 200% is applied. This results in an approximately 30% increase in both Q and $\Delta_z T$, introduced in eqn. 3.4, across the channel. An increase in Q lowers the background $\overline{p\text{CO}_2}$ of the system through increased stratification. However, a similar increase in both Q and $\Delta_z T$ results in a small net change in the already small contribution of $\mathfrak{F}(Q)$ so that the most significant contribution of the stratification $\Delta_z T$ is in the $\overline{\tau_{SO}}^{1/2}$ term, coupled with an increase in stratification is a decrease in mixed layer depth h . In the surface heating experiments, this ranges between 10m in the mid-latitudes to 90m just North of the channel and will increase the wind stress response. The net change in the $\overline{p\text{CO}_2}$ sensitive to wind stress will therefore be determined by the relative changes in h and $\Delta_z T$.

Figure 3.13a shows the change in Southern Ocean $\overline{p\text{CO}_2}$ relative to the control state, $\Delta\overline{p\text{CO}_2}$, for wind stress perturbations with and without a change in radiative forcing. Table 3.13b shows the area-mean of the background parameters C_h , h , $\Delta_z T$ and K in each set of experiments. Both the MITgcm experiments and the scaling show a damped response in $\Delta\overline{p\text{CO}_2}$ to the wind stress changes. The increase in stratification $\Delta_z T$ therefore dominates the change in sensitivity over the mixed layer depth h . It is also worth noting that an increase in wind stress may remove some of the radiative forcing-induced stratification through increased vertical mixing, though this is not apparent in our simulations.

¹⁰Details of the temperature forcing experiments will be discussed in chapter 5.



(a)

Variable	No / 1.3×T warming
$\overline{C_h}$ (mmol/m ³)	2170 / 2164
$\overline{h}$ (m)	300 / 265
$\overline{\Delta_z T}$ (°C)	4.7 / 6.1
$\overline{K}$ (cm/hr)	42 / 42

(b)

Figure 3.13: Summary of the response of the Southern Ocean carbon system to radiative forcing: a) $\Delta p\overline{CO_2}$ as a function of $\Delta\overline{\tau_{SO}}$, with no change in radiative forcing (red) and a 30% increase (green). The squares are MITgcm data and the solid lines are the calculations made using the scaling equation. b) Table of area mean values of parameters C_h , h , Δ_z and K in the Southern Ocean wind stress only and wind stress with warming perturbations experiments.

We have verified that the scaling does predict the correct direction in the shift of $\Delta p\overline{CO_2}$ in response to warming, so we can now apply the scaling to fully coupled global climate models. Under anthropogenic carbon-induced climate change, we expect Southern Ocean stratification to increase along with the rest of ocean. Using the scaling, we can calculate the system's sensitivity to Southern Ocean wind stress under projected climate change conditions using data from global coupled models from the CMIP5 archive. Since the scaling is derived for an equilibrated system, we use $2\times CO_2$ experiments. After 140 years of constant forcing, the $2\times CO_2$ experiments are the closest approximation to an equilibrated coupled GCM available. To illustrate the impact of stratification on the climate feedback shown in Fig. 3.9a, we repeat the calculation, but update the parameters C_h , h and $\Delta_z T$ in the scaling with values taken as the average over the last 10 years of the $2\times CO_2$ experiments for two CMIP5 models, HadGEM2-ES and IPSL-CM5 (we include another CMIP5 model, IPSL-CM5, to illustrate the variety in Southern Ocean representations within CMIP5 models).

The results of the calculation are shown in Fig. 3.14. For HadGEM2-ES, the sensitivity of pCO_2^a to wind stress is reduced in the $2\times CO_2$ experiment compared to the control experiment. However, IPSL-CM5 shows an increase in sensitivity in the $2\times CO_2$ experiment compared to the control experiment. To understand the different response of the climate models, we have to look at each model's response to warming. Figure 3.15 shows the spatial patterns of the mean state pre-industrial air-sea carbon flux (panels a and c in Fig. 3.15) as well as the spatial pattern of the change in the stratification $\Delta_z T$ at the end of the $2\times CO_2$ experiment in the Southern Ocean (panels b and d in Fig. 3.15). There is an overall increase in stratification in HadGEM2-ES. Comparing the pattern of the change in $\Delta_z T$ to the air-sea carbon flux in Fig. 3.15 a, we see that in the key areas of carbon outgassing, the stratification increases. While there are patches in the Pacific and Atlantic sectors of the Southern Ocean where surface cooling causes a reduction in stratification, they are small compared to the large scale pattern of increases in $\Delta_z T$.

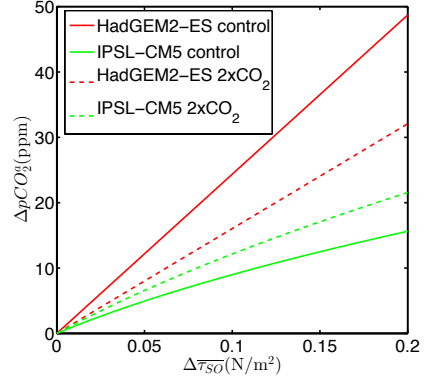


Figure 3.14: ΔpCO_2^a as a function of Southern Ocean mean wind stress between $60^\circ - 45^\circ S$ for CMIP5 models as calculated by the scaling using values for the parameters $\overline{C_h/h}$, $\overline{h}$ and $\overline{\Delta_z T}$ from the pre-industrial control runs (solid lines) and the last 10 years of $2\times CO_2$ experiments (dashed lines).

In IPSL-CM5 on the other hand, there is predominantly cooling in the Southern Ocean (which may be attributed to a latitudinal shift in the wind stress patterns). Figure 3.15 d shows that $\Delta_z T$ decreases in areas of carbon outgassing as a result of a $2\times CO_2$ perturbation which causes a net increase in the sensitivity of $\overline{pCO_2^a}$ to $\overline{\tau_{SO}}$. The Southern Ocean temperature has decreased over the past 30 years [Majkut et al., 2014], therefore indicating that the real-world sensitivity of $\overline{pCO_2^a}$ to $\overline{\tau_{SO}}$ so might also be increasing, similarly to IPSL-CM5.

One might expect a substantial increase in C_h at $2\times CO_2$ over the pre-industrial state but results show that the change in C_h is only 5% on average. The increase in anthropogenic CO_2 results in changes in the chemistry and solubility of carbon that cause pCO_2^a to follow pCO_2^o without increasing C_h substantially. The percentage increase in

stratification is much larger on the other hand and can reach 300% in some models and is therefore the dominant variable in setting the ocean $p\text{CO}_2$ sensitivity.

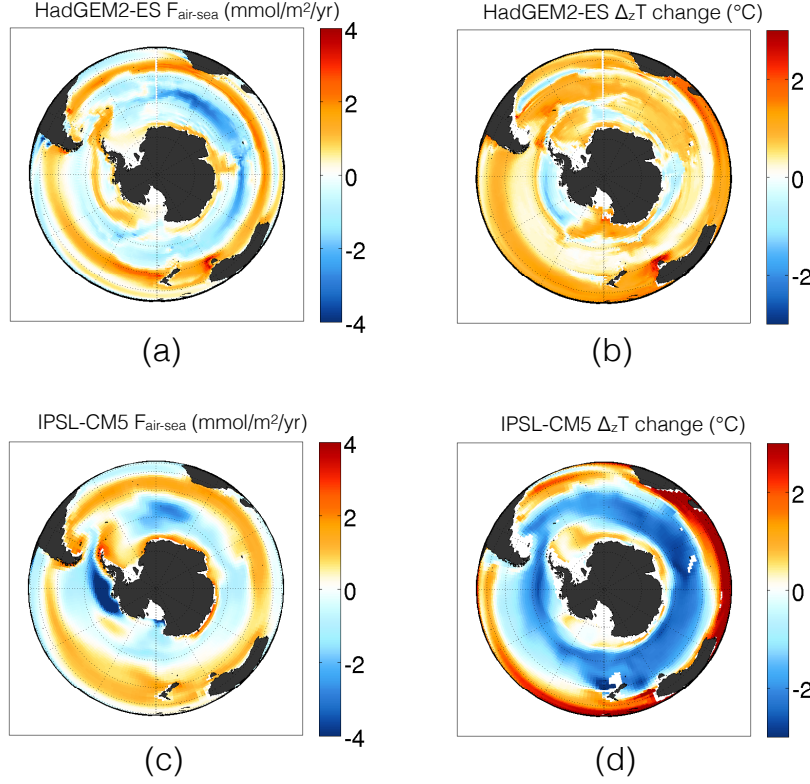


Figure 3.15: Southern Ocean air-sea carbon flux in $\text{mol}/\text{m}^2/\text{year}$ from control runs of A) HadGEM2-ES and C) IPSL-CM5. Change in vertical temperature difference $\Delta_z T$, in $^\circ\text{C}$, between the mixed layer and 1000m depth in B) HadGEM2-ES and D) IPSL-CM5 in the last 10 years of a $2\times\text{CO}_2$ relative to the control run.

3.3 Summary

A mixed layer carbon budget is used to derive an equilibrium equation that balances the advective transport of carbon with the air-sea flux in the Southern Ocean. The diagnostic equation relates perturbations in wind stress to the area mean partial pressure of carbon at ocean's surface $\overline{p\text{CO}_2}$. It is found that $\overline{p\text{CO}_2}$ varies as the square root of the mean wind stress in the Southern Ocean, $\overline{\tau_{SO}}$. The relationship was verified using wind stress perturbation experiments performed with an idealised coarse-resolution MITgcm model set-up representing the Atlantic Ocean with a Southern Ocean-like channel. The square root dependency of $\overline{p\text{CO}_2}$ on $\overline{\tau_{SO}}$ highlights the importance of turbulent entrainment at the bottom of the mixed layer as a result of increased wind stress, rather than any other large scale circulation changes and provides new insight to the Southern Ocean carbon response to wind stress. Using the expression for $\overline{p\text{CO}_2}$, we expressed the atmospheric

partial pressure of CO_2 , $p\text{CO}_2^a$ in terms of the Southern Ocean wind stress. Applying the results to the CMIP5 archive, we find that the parameters h , $\Delta_z T$ and K vary by up to 300% between models, resulting in a similar range of $\Delta p\text{CO}_2^a$ sensitivity to wind stress: 8-25 ppm/(Nm⁻²).

Using experiments with both wind stress perturbations and an atmospheric impulse of CO_2 to represent anthropogenic emissions, we show that the cumulative Southern Ocean carbon uptake also varies as the square root of the area mean wind stress. The mean annual carbon uptake in the Southern Ocean is found to be reduced by 0.04 mol/m²/year for every 0.012 N/m² increase in mean wind stress between 60° – 40°S.

The results presented are calculated using a pre-industrial sensitivity of $\overline{p\text{CO}_2^a}$ on $\overline{\tau_{SO}}$, but the scaling predicts that the sensitivity depends on $\overline{(C_h)} \frac{1}{h\Delta_z T}$, where C_h is the mixed layer carbon concentration, h is the mixed layer depth and $\Delta_z T$ is the temperature difference between the mixed layer and at depth (1000m). We use MITgcm radiative forcing and wind stress perturbation experiments to show that with increased stratification due to radiative forcing, the $\overline{p\text{CO}_2^a}$ response to $\overline{\tau_{SO}}$ is reduced. The sensitivity is dominated by the stratification $\Delta_z T$ such that estimates made using the scaling capture the reduction in sensitivity. While we have so far assumed C_h , h and $\Delta_z T$ do not depend on wind stress, further study is required to determine the exact dependence of these parameters on wind stress.

In CMIP5 models, the mean stratification induced by a 2× CO_2 perturbation (through surface warming, freshwater or mechanical forcing) can either increase or decrease the sensitivity of $\Delta p\text{CO}_2^a$ on $\overline{\tau_{SO}}$, depending on whether the Southern Ocean warms or cools in response to anthropogenic CO_2 forcing. Previous studies such as Lovenduski and Ito [2009] also look at the effects of wind stress and stratification separately. However, Lovenduski and Ito [2009] suggest that when they apply both wind stress and buoyancy perturbations at once, the result is a linear superposition of the two contributions. However, we show that an increase in stratification reduces the sensitivity of the system to wind stress so that the two separate contributions do not simply add linearly.

One caveat is that the MITgcm model set-up uses a parametrization to represent eddy transports. However, since most CMIP5 models use the same parametrization, our

results and scaling are appropriate for assessing wind stress feedbacks for the majority of the CMIP5 archive. Nonetheless, studies such as Munday et al. [2014] and Jones et al. [2011] show that resolved eddies reduce the Southern Ocean carbon response to wind stress perturbations through eddy compensation, as a result of the effect of eddies on the residual circulation [Hallberg and Gnanadesikan, 2001, Shakespeare and Hogg, 2012]. While further study is needed to determine the exact impact of eddies on the feedback, we estimate the contribution of eddies towards Southern Ocean pCO_2^o change by performing simulations with an increased eddy diffusivity and estimates of bolus velocities from eddy-permitting simulations performed by Munday et al. [2013]. The net effect on the meridional advection term towards $\overline{pCO_2^o}$ is limited to -5ppm for changes up to 0.4 N/m^2 in $\overline{\tau_{SO}}$. However, the presence of stronger eddies affects the background stratification. We find that a 30% increase in the eddy diffusivity parameter κ_{GM} increases area-mean $\Delta_z T$ by 0.2° and the mixed layer depth h by 5 m. The net result is a reduction in the sensitivity of Southern Ocean pCO_2^o to wind stress by 20%, shown in Fig. 3.16 (cyan). We can apply the scaling equation with the perturbed values of $\Delta_z T$ and h from the $1.3 \times \kappa_{GM}$ experiments and find that it predicts the correct shift in the sensitivity. The scaling can therefore still be used with higher-resolution models or models with variable κ_{GM} as long as the parameters $\Delta_z T$ and h are known.

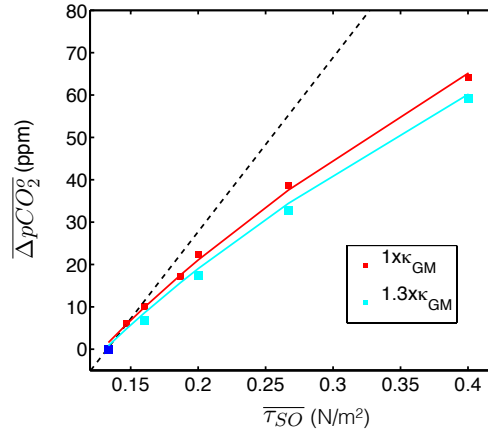


Figure 3.16: Change in the MITgcm Southern Ocean $\overline{pCO_2^o}$, as a function of mean channel wind stress. Squares denote MITgcm data after 400 years for the control run (dark blue), Southern Ocean wind perturbation (red) and wind perturbation with 30% stronger GM coefficient (cyan) experiments. The red and cyan lines are the calculations using the scaling and the dashed black line is a linear trend around the control state for the standard wind perturbation experiments.

The non-local climate feedback of Southern Ocean winds

The response of the Southern Ocean carbon sink to Southern wind stress has been well-studied [e.g., Le Quere et al., 2007, Lovenduski et al., 2008, Lovenduski and Ito, 2009, Ito et al., 2010, Lauderdale et al., 2013, Munday et al., 2014]. However, there is little understanding of the climate impacts of dynamical and biogeochemical changes induced by the Southern Ocean wind-stress outside of the Southern Ocean. Studies such as Lauderdale et al. [2013] examine the basin-scale response of perturbations in Southern Ocean wind stress on the dynamics and distribution of carbon, without identifying the local vs. non-local contributions of the different processes on atmospheric $p\text{CO}_2$. Similarly, studies such as Anderson et al. [2002], Sarmiento et al. [2004] tie Southern Ocean upwelling to the nutrient supply North of the Southern Ocean, but without further implications for the climate including atmospheric $p\text{CO}_2$. We aim to quantify the impact of changes in ocean circulation and biogeochemistry occurring North of the Southern Ocean due to perturbations in Southern Ocean wind stress on atmospheric $p\text{CO}_2$. We find that increased Southern Ocean wind stress reduces the capacity of the North Atlantic sub-polar gyre to absorb atmospheric CO_2 by affecting the biogeochemistry of the sub-polar gyre water upstream, before it is upwelled in the Northern high latitudes.

In the Atlantic, the major oceanic sink of atmospheric carbon (both natural and anthropogenic) is the sub-polar gyre, which plays a key role in the meridional overturning circulation and uptake of carbon [Group, 1998, Dickson et al., 2008, Kieke and Yashayaev, 2014]. Therefore any changes induced in ocean $p\text{CO}_2$ in the sub-polar gyre will greatly impact the atmospheric carbon concentrations by affecting oceanic CO_2 uptake. The biogeochemical properties of the sub-polar gyre are determined by the water advected from the Tropics and Southern Ocean. Changes in Southern Ocean wind stress therefore affect carbon and nutrient transport pathways throughout the ocean basins and can lead to ocean $p\text{CO}_2$ changes in remote regions such as the North Atlantic subpolar region. In section 4.1, we examine changes in the Atlantic carbon and nutrient transport pathways and the overall climate-carbon feedback due to Southern Ocean wind stress. To put the

results into perspective, section 4.2 compares the climate-carbon feedback in the North Atlantic sub-polar gyre induced by local wind stress changes to the non-local feedback induced by Southern Ocean winds.

4.1 Southern Ocean influence on North Atlantic sub-polar gyre chemistry

North Atlantic $p\text{CO}_2$ and atmospheric $p\text{CO}_2$ are closely linked. To illustrate the role of the North Atlantic sub-polar gyre ocean $p\text{CO}_2$ on atmospheric $p\text{CO}_2$, Fig. 4.1 shows time series of area-averaged¹ anomalies of ocean $p\text{CO}_2$, $\Delta\overline{p\text{CO}_2^o}$, for the Southern Ocean ($40^\circ - 60^\circ\text{S}$) and the North Atlantic sub-polar gyre ($40^\circ - 60^\circ\text{N}$) and anomalies of atmospheric $p\text{CO}_2$, $\Delta p\text{CO}_2^a$, in the $2 \times \tau_{SO}$ wind stress perturbation experiment with the MITgcm.² As discussed in section 3.1, the initial adjustment of the Southern Ocean $p\text{CO}_2$ occurs within 50 – 100 years of the wind-stress perturbation. The atmospheric signal shows an initial response which appears to correspond to the timescale of the local response of the Southern Ocean to the increase in wind stress. The short-term response is followed by a long-term centennial adjustment in $p\text{CO}_2^a$ due to other adjustments of the ocean carbon system, which we refer to as the non-local effect. The long-term adjustment of the atmospheric $p\text{CO}_2$ shows a timescale similar to the North Atlantic sub-polar gyre $p\text{CO}_2$ response to the non-local wind stress change, more so than other regions (not shown). After 400 years, atmospheric $p\text{CO}_2$ is mostly equilibrated so stays constant for longer integrations.

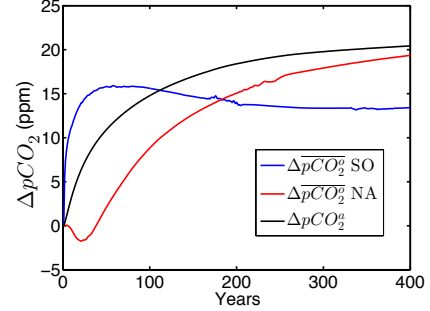


Figure 4.1: Time evolution of $\Delta p\text{CO}_2^a$ (black), North Atlantic $\Delta\overline{p\text{CO}_2^o}$ (red) and Southern Ocean $\Delta\overline{p\text{CO}_2^o}$ (blue) in a $2 \times \tau_{SO}$.

In the following sections, we aim to show that increased Southern Ocean winds affect the North Atlantic CO_2 uptake through a novel mechanism: increased Southern Ocean wind stress enhances nutrient transport to the Tropics, thereby increasing tropical biological productivity which in turn affects the chemistry of underlying sub-thermocline waters. The tropical sub-thermocline waters are subsequently advected to the Atlantic sub-polar

¹The overbar denotes an average over the region of interest defined in parenthesis.

²Anomalies are defined as $\Delta X = X - \langle X \rangle$, where $\langle \dots \rangle$ indicates the temporal mean of the control run.

gyre by the MOC, reducing the capacity of the sub-polar gyre to absorb atmospheric CO₂. The following sections will outline the key steps in the mechanism:

1. The impact of Southern Ocean winds on nutrient transport will be discussed through the concept of a nutrient stream, which is a meridional transport pathway in the ocean along which the majority of nutrients are transported from the Southern Ocean to the Northern hemisphere.
2. Surface biology affects the chemistry of waters below the thermocline through vertical transfer of DIC and Alkalinity which both determine the Revelle Buffer Factor R . The Revelle Buffer Factor measures the capacity of water parcels to absorb atmospheric CO₂.
3. The Nutrient stream is modified by biogeochemical signal advected out of the tropical shadow zone.
4. Together, the interaction of the nutrient stream, the shadow zone dynamics and North Atlantic biology result in the observed decrease in the North Atlantic sub-polar gyre CO₂ uptake.

4.1.1 Nutrient streams

The increase in the Southern Ocean residual circulation due to increased wind stress causes an increase in upwelling of deep carbon- and nutrient-rich waters as described in section 3.1.1. In the Southern Ocean, surface biology is inefficient due to strong light and iron limitation [Boyd et al., 2000, Coale et al., 2004, DeVries et al., 2012] so the additional upwelled-nutrients are mostly unused. Instead, the nutrients are transported northwards and into the basin interior following the sub-surface currents in what are commonly referred to as nutrient streams [Pelegri and Csanady, 1991, Pelegri et al., 1996, 2006, Williams et al., 2006, 2011].

A nutrient stream is an advective pathway, through which nutrients are transported meridionally between oceans. In particular, they are important for transporting nutrients from regions of upwelling and nutrient excess (i.e. Southern Ocean) to nutrient-poor regions such as the equator and the Northern high latitudes (see for example Fig. 4.2a for the MITgcm control run nutrient distribution). Ekman transport drives nutrients northwards and out the Southern Ocean [Williams and Follows, 1998], where it is subducted

below the mixed layer during formation of Antarctic Intermediate Water, where the cold northward advected water masses are subducted below warmer southwards advected waters from the sub-tropical gyre. Transport is then predominantly isopycnal, with nutrients gradually re-inducted into the mixed layer as isopycnals shoal towards the surface in the North Atlantic sub-polar region [Williams et al., 2006].

North of the Southern Ocean where biology is far more efficient than in the Southern Ocean itself, the nutrient stream causes an increase in macro-nutrients³ limited biological productivity [Sarmiento et al., 2004]. The nutrient stream $S_P(x, y, z, t)$ is defined here as the meridional transport of Phosphorus, in the form of Phosphate PO_4^{3-} , by the residual velocity field v and is given by

$$S_P(x, y, z, t) = [PO_4^{3-}] v. \quad (4.1)$$

In order to understand the biogeochemical response to the wind stress changes, we also define the zonal nutrient transport $S_{Px} = [PO_4^{3-}] u$, where u is the residual zonal velocity (we will refrain from calling this a nutrient stream for consistency with the literature).

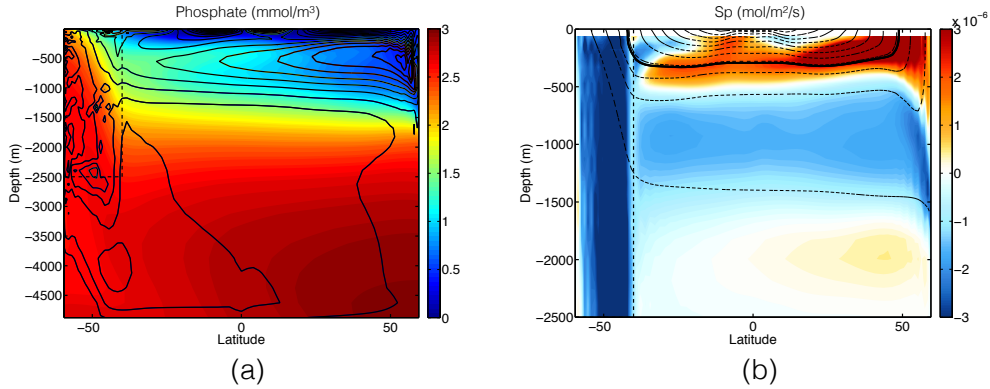


Figure 4.2: Zonal mean for the control run of a) PO_4^{3-} overlain with stream function (black contours) and b) S_P overlain with potential density contours (dashed black lines). The thicker solid black line shows the density surface $\sigma = 26 \text{ kg/m}^3$.

Physical transport in the ocean interior is mostly adiabatic, so the location of the nutrient stream lies along surfaces of constant density (see Fig. 4.2b), although the strength can vary depending on the biological flux of nutrients [Pelegri et al., 1996]. In the case of the MITgcm experimental set-up, the maximum transport occurs along the constant density surface $\sigma = 26 \text{ kg/m}^3$, shown in Fig. 4.3a for the control run, North of the

³Mainly Nitrogen and Phosphorous.

Southern channel. Therefore we interpolate the meridional transport of phosphate onto the density surface $\sigma = 26 \text{ kg/m}^3$. Figure 4.3 also shows the zonal nutrient transport S_{Px} on the density surface $\sigma = 26 \text{ kg/m}^3$ (panel b), the steady state concentration of PO_4^{3-} along $\sigma = 26 \text{ kg/m}^3$ (panel c) and the biological productivity integrated over the mixed layer, $[F_{bio}]_h$ (panel d). The spatial pattern of S_P shows the nutrient stream, the region of positive S_P , originating just North of the Southern channel where intermediate mode water is formed in region (1). In region (2), there is a recirculation due to the imprint of the western boundary current of the Southern subtropical gyre. The nutrient stream follows the northwards MOC transport (3), increases in strength until it reaches the sub-polar gyre in region (4), where the nutrients are entrained into seasonal thermocline through mixing. The zonal nutrient transport S_{Px} shows a westward transport along the equatorial undercurrent and an eastward transport along equatorial countercurrents to the North and South of the central westward current (region 5).

The effect of increased Southern Ocean winds on the nutrient transport is shown in Figure 4.3 (bottom row) for the $2 \times \tau_{SO}$ perturbation experiment. Due to the increase in Southern Ocean wind-induced vertical mixing, there is a general increase in concentrations of PO_4^{3-} along the $\sigma = 26 \text{ kg/m}^3$ surface in the Southern hemisphere (panel g). The wind stress perturbation also causes an increase in the MOC and the meridional velocity v along the $\sigma = 26 \text{ kg/m}^3$ surface. In combination with the increased concentrations of PO_4^{3-} , the result is an increase in the nutrient stream (panel e). The eastward Southern equatorial countercurrent intersects the nutrient stream, resulting to an increase in eastward nutrient transport into the shadow zone (by 102% in the $2 \times \tau_{SO}$ experiment, see region 6). The increase in eastward nutrient transport is also responsible for delivering phosphate into the nutrient-limited shadow zone (region 6). Subsequently, mixed layer biological productivity is strongly enhanced North of the Southern channel and in the eastern Tropics where nutrients are upwelled due to wind stress divergence (panel h). For the $2 \times \tau_{SO}$ experiment, the increase in biological productivity is less than 10% in the Southern channel, between 20% and 40% in the sub-polar gyre, but more than 200% between 20°S and 20°N.

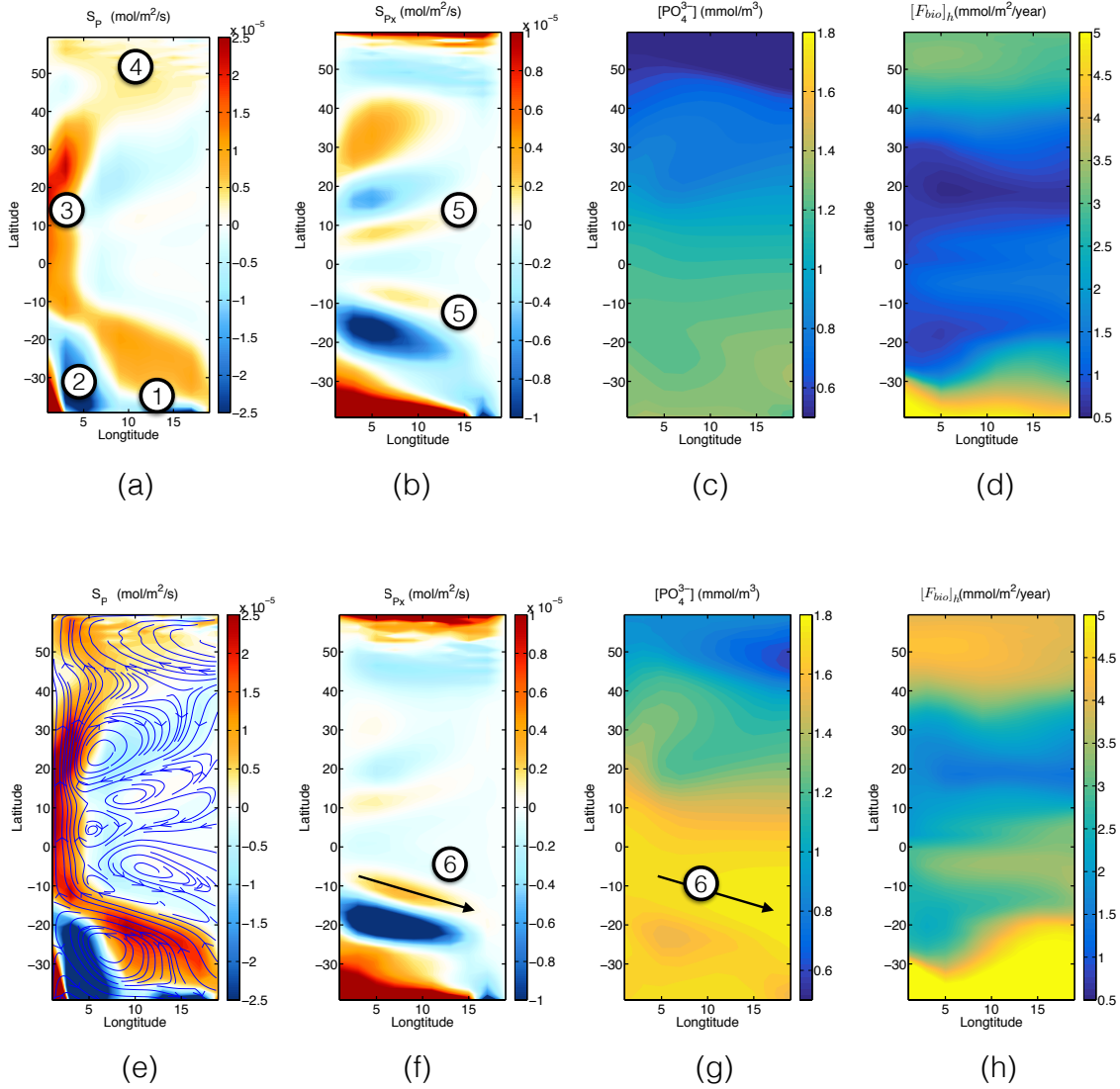


Figure 4.3: The nutrient stream S_p (a, d), the zonal nutrient transport S_{Px} (e, d), phosphorous (c, f) on the $\sigma = 26 \text{ kg/m}^3$ surface and the mixed layer biological productivity $[F_{bio}]_h$ (d, g). Top row is for the control run and bottom row is for the $2 \times \tau_{SO}$ perturbation experiment. Streamlines along the $\sigma = 26 \text{ kg/m}^3$ surface are shown in blue in panel (e). The figure only shows the fields North of the Southern channel for clarity.

The surface biological productivity in the eastern Tropics is sustained mostly through the upwelling of remineralized nutrients within the shadow zone itself, causing local cycling of nutrients and coined ‘nutrient trapping’ [Najjar, 1990, Aumont et al., 1999]. The build up of nutrients in the shadow zone also explains the equatorial sub-surface maximum in regenerated carbon observed in most complex ocean models [Rhoads and Morse, 1971, Stramma et al., 2009]. Such a maximum in nutrients and regenerated carbon is also seen

in observations through a corresponding minimum in oxygen [Karstensen et al., 2008].

In the Northern sub-polar gyre, the response in productivity is not as strong as in the Tropics. The relatively weak response in the sub-polar gyre is partly due to high-latitude light and iron limitation [Nielsdottir et al., 2009] and partly due to the nutrient stream depletion of nutrients by the time it reaches the Northern end of the basin. Most of the increase in biology is centred around the eastern Tropics. At depth, there is a large increase in regenerated carbon underneath the thermocline in the eastern Tropics, as shown in Fig. 3.5b. The change in regenerated carbon due to remineralisation, δC_{reg} , is governed by equation 3.2 in steady state in the ocean interior: $v \frac{\delta C_{reg}}{\delta y} = r_{C:P} \mathcal{F}_{PO_4}$.⁴ The only significant increase in C_{reg} along the $\sigma = 26 \text{ kg/m}^3$ surface occurs in the eastern Tropics well outside the nutrient stream in the Southern hemisphere, and downstream in the Northern hemisphere nutrient stream, as shown in Fig. 4.4. Within the Southern hemisphere section of the nutrient stream, both remineralization $r_{C:P} \mathcal{F}_{PO_4}$ and meridional velocity v increase by similar percentages resulting in a small net change in C_{reg} .

4.1.2 Surface biology and the Revelle Buffer Factor

An increase in equatorial biology can act as a negative feedback on atmospheric pCO_2^a by drawing more carbon from the mixed layer, reducing ocean surface pCO_2 and therefore increasing ocean CO_2 uptake. Conversely, it can also act as a positive feedback by affecting the chemistry of sub-thermocline waters by altering the Revelle Buffer Factor of the underlying intermediate water. The buffer factor R quantifies how much additional atmospheric carbon the ocean can absorb, therefore making any changes in R relevant for climate. For example, water with a positive R -anomaly that upwells in

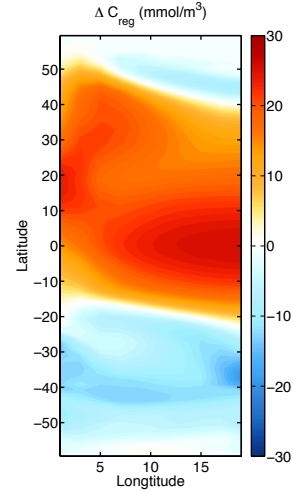


Figure 4.4: Change in C_{reg} along the $\sigma = 26 \text{ kg/m}^3$ density surface after 400 years of a $2 \times \tau_{SO}$ perturbation.

⁴In the ocean interior where the dominant forcing on carbon is remineralization, the carbon cycle components C_{pre} and C_{reg} can be treated as independent, since they only interact with one another at the surface by definition [Williams and Follows, 2011].

the Northern Hemisphere sub-polar gyre will increase the local $p\text{CO}_2^o$, therefore decreasing the carbon sink. The Revelle buffer factor is generally defined as (see section 1.2.4):

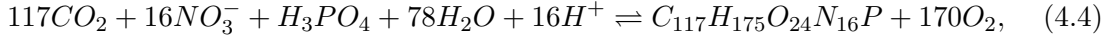
$$R = \frac{\delta p\text{CO}_2}{p\text{CO}_2} / \frac{\delta \text{DIC}}{\text{DIC}}. \quad (4.2)$$

Equation 4.2 can be simplified by: 1) invoking conservation of charge, 2) recognizing that fractional changes in dissolved CO_2 ($[\text{CO}_2^*]$) are approximately equal and opposite to those of $[\text{CO}_3^{2-}]$, 3) the magnitude of the perturbations in $[\text{CO}_2^*]$ are much smaller than those in $[\text{CO}_3^{2-}]$. We can then write the buffer factor as (see Williams and Follows [2011]):

$$R = \frac{\text{DIC}}{[\text{CO}_3^{2-}]} \approx \frac{\text{DIC}}{\text{Alk} - \text{DIC}}, \quad (4.3)$$

where Alk is Alkalinity, since $[\text{CO}_3^{2-}] \approx \text{Alk} - \text{DIC}$.

Surface biology utilises DIC and Alkalinity at different rates. Biology therefore leaving an imprint on the Buffer Factor both at the surface (in the mixed layer) and at depth. To understand the effect of biology on the Buffer Factor, let us first consider the influence of mixed layer biology on the mixed layer Buffer Factor. Production of soft matter follows the Redfield equation (using recent circulation averaged stoichiometry) [Redfield, 1934, Anderson and Sarmiento, 1994, Weber and Deutsch, 2012]:



where the forward and backward reactions represent production and remineralization, respectively. Using the definition for total Alkalinity, $\text{Alk} = [\text{HCO}_3^-] + 2[\text{CO}_3^{2-}] + [\text{OH}^-] - [\text{H}^+] + [\text{B}(\text{OH})_4^-]$,⁵ production of soft matter reduces DIC by 117 moles for every mol of phosphate and increases Alkalinity by 16 moles. The formation or dissolution of calcite follows the equation:



The formation of 1 mol CaCO_3 requires of 1 mol of CO_3^{2-} which will reduce DIC by 1 mol and Alkalinity by 2 mol. Surface biology therefore changes the local concentrations of DIC and Alk which will change the local Buffer Factor R , depending on the relative export of soft matter and calcite. The relative export of soft matter to calcite, r , is known as the “rain ratio.” Climatological studies of basin-wide data such as Sarmiento et al. [2002], Jin

⁵The conserved ions NO_3^- and HPO_4^{2-} are not considered here since their concentrations are large enough such that small variations do not affect alkalinity [Sarmiento and Gruber, 2006].

et al. [2006] give a rain ratio between 10-20, the former is used in the MITgcm experiments conducted here. Higher estimates based on local and seasonal data can be obtained, for example Emerson et al. [2011] estimate local rain ratios down to 2. Even at the lowest local estimates of the rain ratio (2), the net result is that the mixed layer Revelle Buffer Factor R is *always* reduced upon the formation of organic carbon and calcite: surface biology reduces DIC and increases Alkalinity in surface waters of the open ocean.⁶ In the Southern Ocean wind stress experiments, biology is increased in the Southern channel, the Tropics and the North Atlantic sub-polar resulting in a local reduction in R in each region.

Surface biology also affects the vertical transport of DIC and Alkalinity in the ocean. Therefore, we will consider how surface biology affects the Buffer Factor R at depth. The soft matter that is produced in the ocean mixed layer falls vertically through the water column until it is remineralized back into DIC, in the form of C_{soft} . The remineralisation of soft matter follows the reverse of eqn. 4.4, therefore increasing DIC by 117 mols and reducing Alkalinity by 16 for every mol of phosphate. Similarly, calcite shells produced in the mixed layer fall vertically through the water column until they are dissolved into DIC in the form of C_{carb} according to the reverse of eqn. 4.5. Dissolution of 1 mol of calcite therefore increases DIC by 1 mol and Alkalinity by 2 mol. For all observed values of the rain ratio r , the net effect is an increase in DIC and a reduction in Alk at depth. In regions of biological activity, the underlying waters are therefore enhanced in R . If these water parcels are then advected across the ocean and upwelled to the surface, they will increase R in the region of upwelling. Since R measures the capacity of the water to absorb atmospheric CO₂, water parcels containing an increased amount of remineralized organic matter will therefore be upwelled with an elevated buffer factor R and will be less capable of absorbing atmospheric CO₂ anomalies.

In our MITgcm experiments, the strongest increase in the Buffer Factor R at depth occurs below the thermocline in the equatorial shadow zone, shown in Fig. 4.5a after 50 years and Fig. 4.5b after 400 years. The evolution of R shows the following pattern: an initial increase in R in the Southern Channel due to upwelling which is reduced after 400 years, a reduction in R between 1000m and 2500m depth, an increase in R in AABW and a region of strong increase in the Tropics between 200m and 700m depth. Below the

⁶As a reminder, the rain ration r used in the MITgcm is 20.

thermocline, the major forcing in R is biology: Fig. 4.5c-f show the temporal evolution of the buffer factor, R , anomaly along the $\sigma = 26 \text{ kg/m}^3$ density surface. After 50 years, there is an increase in R in the Southern channel associated with increased upwelling of DIC-rich waters. A positive R anomaly occurs in the Tropics, North of the edge of the nutrient stream, which originates from the eastern boundary due to the increased biological activity. After another 100 years, the anomaly in R in the Southern channel has decreased while the tropical anomaly signal intensifies.

An increase in R is generated in the eastern Tropics (region 1 in Fig. 4.5c) and transported westward within the shadow zone, where it is then entrained by the nutrient stream (denoted by the solid arrows in panel f). There is also a recirculation within the shadow zone such that a part of the R -anomaly is entrained South of the equator in region 2 (denoted by the dashed arrows in panel f). Subsequently, the signal follows the nutrient stream northward (region 3) into the sub-tropical gyre where it upwells. Part of the R anomaly generated in the Southern Ocean (the “preformed” signal), denoted by region 4 in Fig. 4.5c, follows the nutrient stream as well, but the signal is between 10% and 30% of that in the equatorial region (the “regenerated” signal). The R -anomaly upwelled in the Southern Ocean is weakened when it is subducted into intermediate waters since the intermediate waters from Southern origin have been ventilated at the surface, where the biology and air-sea fluxes have reduced the upwelling R -anomaly. On the other hand, the waters in the eastern tropical shadow zone are only ventilated in the Northern sub-polar gyre. The gradual increase in transport of the buffer factor anomaly from the equator northwards along the western boundary is also shown in the Hovmuller diagram in Fig. 4.5g.

The positive R anomaly from the shadow zone is therefore advected to the North Atlantic sub-polar gyre. As was discussed earlier, surface biology reduces the Buffer Factor of the local mixed layer. While the surface biology in the sub-polar gyre reduces R , by the time the nutrient stream reaches the sub-polar gyre it is nearly stripped of the nutrient surplus so that the increase in local light and iron limited-biology in the North Atlantic sub-polar gyre is not enough to sufficiently reduce the positive R -anomaly of the upwelled waters.⁷ The net result is therefore an increase in the North Atlantic sub-polar gyre Buffer Factor as a result of increased Southern Ocean wind stress.

⁷There is also a positive R -anomaly in the North-East corner which is associated with the re-subduction of C_{reg} in the form C_{dis} .

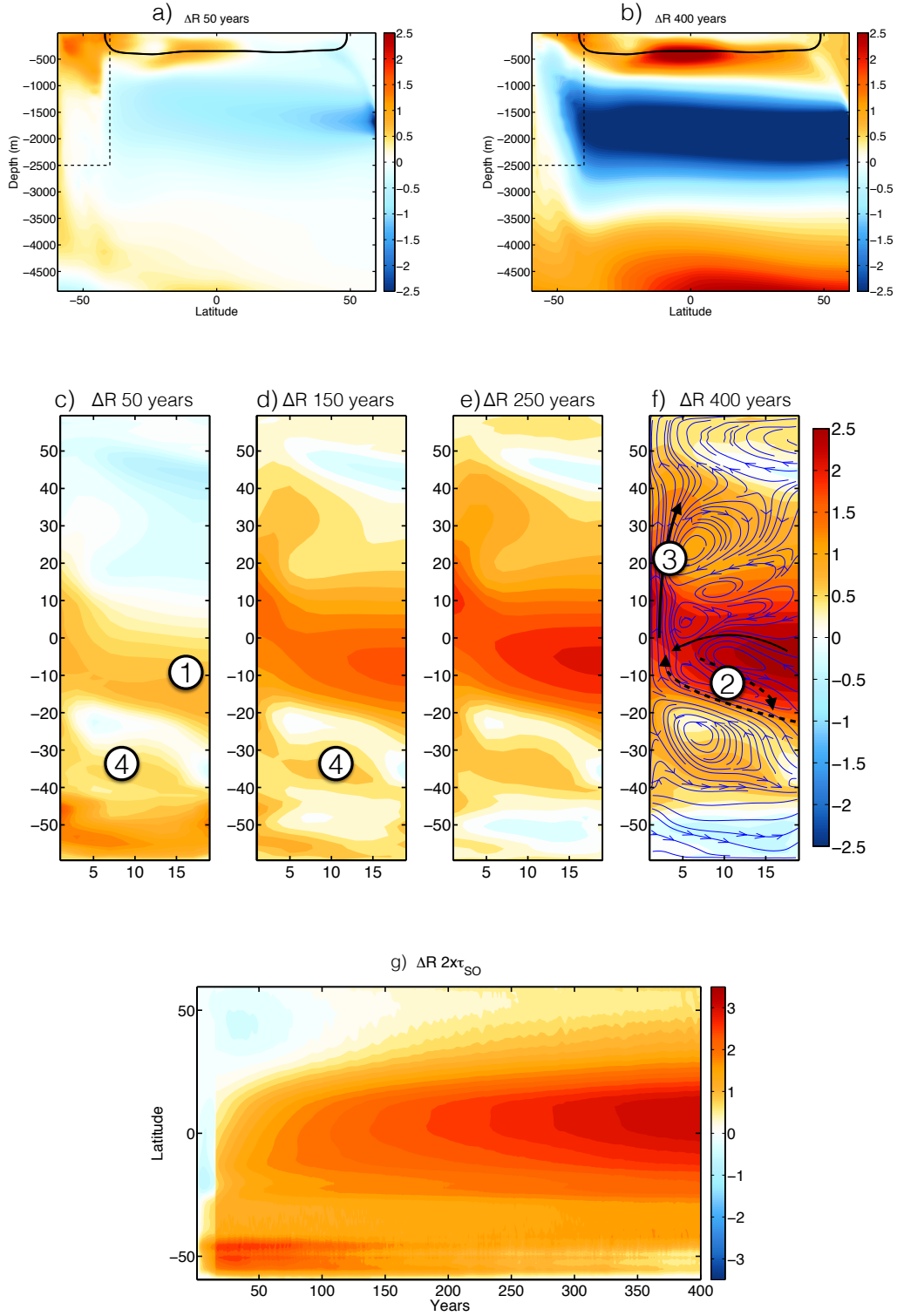


Figure 4.5: Zonally averaged anomaly ($2 \times \tau_{SO}$ - control) of the Revelle Buffer Factor R after a) 50 years and b) 400 years. The solid line shows the depth of the perturbed $\sigma = 26$ kg/m³ surface. Revelle Buffer Factor anomaly ($2 \times \tau_{SO}$ - control) along the perturbed density surface $\sigma = 26$ kg/m³ after (c) 50, (d) 150, (e) 250 and (f) 400 years, where the blue lines show the perturbed streamlines. Panel g) is a Hovmuller diagram of the buffer factor anomaly between 0 – 4°E, with latitude over 400 years.

4.1.3 The North Atlantic sub-polar gyre carbon sink

So far, we have shown that an increase in Southern Ocean wind stress causes an increase in the nutrient stream S_P , resulting in an increase in equatorial mixed layer biological activity. The surface biology increases the buffer factor R in the equatorial shadow zone at depth through increased remineralization. Subsequently, the positive R anomaly is entrained into the nutrient stream and upwelled in the North Atlantic sub-polar gyre. Since R affects the capacity of the ocean to absorb atmospheric CO_2 , the value of R in the Northern sub-polar gyre will affect the air-sea flux of CO_2 . Figure 4.6 shows time series of the evolution of $\overline{p\text{CO}_2^o}$, $\overline{F_{air-sea}}$ and $\overline{R}$ in the Northern sub-polar gyre in the $2 \times \tau_{SO}$ perturbation experiment (where the overline is the spatial average between 40° and 60°N). After several decades, the positive equatorial R -anomaly is upwelled in the sub-polar gyre causing a rise in $\overline{p\text{CO}_2^o}$. The net result is a decline in the air-sea flux to a value smaller than that of the control run.⁸

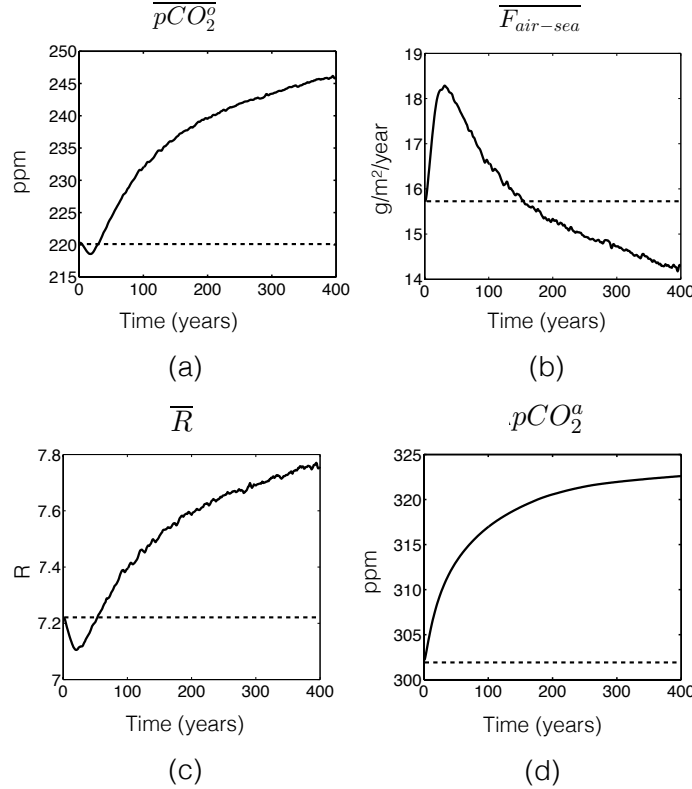


Figure 4.6: Time series of (a) $\overline{p\text{CO}_2^o}$, (b) $\overline{F_{air-sea}}$ and (c) mixed layer-averaged $\overline{R}$ in the North Atlantic sub-polar gyre between $40^\circ - 60^\circ\text{N}$ and (d) atmospheric $p\text{CO}_2$ in the $2 \times \tau_{SO}$ perturbation experiment.

⁸In the initial 10-20 years of the perturbed experiment, convective adjustment occurs at high latitudes which reduces upper ocean DIC and Alkalinity in the sub-polar gyre lowering $\overline{R}$ and $p\text{CO}_2^o$.

The decline of the North Atlantic air-flux of carbon has important implications for the overall climate feedback of atmospheric CO_2 to Southern Ocean wind stress. Since the North Atlantic sub-polar is key sink of atmospheric CO_2 , a decrease in the air-sea flux will result in elevated atmospheric $p\text{CO}_2$ levels. Moreover, the increase in atmospheric $p\text{CO}_2$ due the North Atlantic sub-polar gyre is caused by an increase in equatorial biology. While the increase in surface biology in the tropics results in a local drawdown of atmospheric carbon, the strength of equatorial biology is still relatively weak compared to the high latitudes. However, equatorial biology affects the chemistry of the waters in the North Atlantic, where the strength of the air-sea flux of carbon is key for determining the atmosphere-ocean balance of carbon. The local drawdown of carbon in the tropics by biology is therefore outweighed by the preconditioning of North Atlantic waters. A mechanistic understanding is therefore presented for explaining why an increase in surface biology in one region can lead to an increase in atmospheric $p\text{CO}_2$.

4.1.4 Equatorial biology and the non-local climate feedback of Southern Ocean wind stress

Past studies show that increases in ocean biology lead to drawdowns in atmospheric CO_2 levels. For example, Ito and Follows [2005, 2013] link $p\text{CO}_2^g$ to the efficiency of global nutrient utilisation N^* , showing that $p\text{CO}_2^g$ decreases as the strength of global biology increases. However, we find that an increase in equatorial biology can lead to an increase in atmospheric $p\text{CO}_2$. Should the efficiency of global biological productivity increase, the amount of regenerated carbon stored in the ocean increases at the expense of atmospheric carbon, however the effect of spatial patterns of biological activity can play a vital role in the uptake of atmospheric CO_2 . Many studies such as Sarmiento and Orr [1991], Falkowski et al. [1998], Marinov et al. [2006], Ito and Follows [2013] show that high-latitude biology, particularly in the Southern Ocean, has a strong influence on $p\text{CO}_2^g$. However, little attention has gone to equatorial biology.

Sarmiento and Orr [1991] and Falkowski et al. [1998] show that an increase in high-latitude biology results in a drawdown of $p\text{CO}_2^g$, given no large-scale changes in mechanical or buoyancy forcing. The positive feedback of the equatorial biology on $p\text{CO}_2^g$ in the wind-stress perturbation experiments is therefore in direct competition with the negative feedback of the biology in the North Atlantic sub-polar gyre. We define the strength of

the mixed layer mean biological flux in the North Atlantic sub-polar gyre (averaged over 40°-60°N), $\overline{F_{bio,na}}$, and in the equatorial region (10°S to 10°N), $\overline{F_{bio,eq}}$. Guided by eqns. 7.4 and 2.14, we make the following set of assumptions, which we will attempt to verify later on:

- The flux of biological carbon at any depth z is proportional to the flux of carbon F_{bio} (where $F_{bio} = r_{C:P}\mathcal{F}_{PO4}$) out of the mixed layer [Martin et al., 1987]. We can insert F_{bio} into equation 3.2 as a source term and infer that ΔC_{reg} underneath the mixed layer is proportional to $\overline{F_{bio,eq}}$. For a change in regenerated carbon, ΔC_{reg} , in the eastern tropical shadow zone, the mixed layer buffer factor in the North Atlantic sub-polar gyre R_{na} will increase as ΔC_{reg} increases. Therefore, the North Atlantic buffer factor $\overline{R_{na}}$ is proportional to the equatorial biology $\overline{F_{bio,eq}}$.
- Conversely, surface biological activity in the North Atlantic strips the water of DIC and Alkalinity and acts to reduce the Revelle Buffer Factor in the mixed layer. We find that the North Atlantic buffer is inversely proportional to the North Atlantic biological flux of carbon:

$$\overline{R_{na}} \propto 1/\overline{F_{bio,na}}. \quad (4.6)$$

- By definition of the Revelle Buffer Factor, ocean pCO_2 increases linearly as the buffer factor increases leading to ocean outgassing of CO_2 . An increase in the surface ocean buffer factor therefore leads to a reduction of the air-sea carbon flux into the ocean, causing an increase in atmospheric pCO_2 .

As a result, the Buffer Factor in the North Atlantic is set by the ratio of the equatorial biological flux over the sub-polar biological flux, which in turn will lead to a corresponding change in atmospheric pCO_2 , leading to a positive feedback:

$$pCO_2^a \propto \overline{R_{na}} \propto \frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}. \quad (4.7)$$

The ratio $\frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$ is 0.5 in the MITgcm control integration. The feedback loop is summarised in Fig. 4.7.

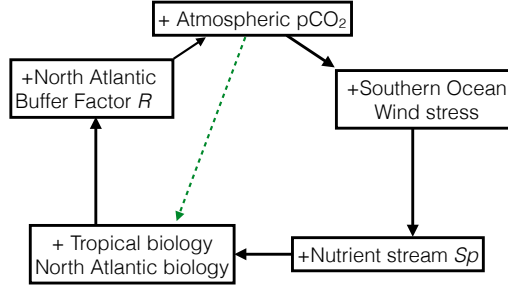


Figure 4.7: Positive feedback loop of the non-local atmospheric CO_2 -induced Southern Ocean wind stress changes (black arrows). The green arrow shows that a similar feedback can be achieved by changes in biological productivity due to forcing other than Southern Ocean wind stress such as iron deposition or stratification changes.

To explore the relationships hypothesised above and effect of biology on $\overline{R_{na}}$ and $p\text{CO}_2^a$, a further series of 11 experiments were run perturbing parameters in the biogeochemical model in MITgcm, namely the nutrient limitation half-saturation constant K_N (5 experiments) and the light limitation half-saturation constant I_k (6 experiments), details of which are in the appendix (section 7.7). The effect of perturbing the parameters is to increase or decrease biological activity by an amount which depends on the local light or nutrient limitation, without affecting the circulation so that the changes seen in $\overline{F_{bio,eq}}$ and $\overline{F_{bio,na}}$ are purely biogeochemical. Increasing K_N causes areas that were not limited in macro-nutrients (e.g. high latitudes) to become limited, effectively increasing $\frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$. Increasing I_k has the same effect on areas that are usually not light limited (e.g. the Tropics), effectively reducing $\frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$.

The nutrient and light half-saturation perturbation experiments can be used to illustrate the feedback loop shown by the green arrow in Fig. 4.7, such that the increase in atmospheric $p\text{CO}_2$ can also be caused by perturbations in the biology due to factors others than wind stress. Therefore the proposed feedback mechanism is not solely dependent on DIC transport due to changes in Southern Ocean residual circulation but applies generally. The mechanism is also tested in wind stress perturbation experiments with a simultaneous increase in radiative forcing (a step $1.2 \times T$ base warming perturbation, as described in section 2.3.4).

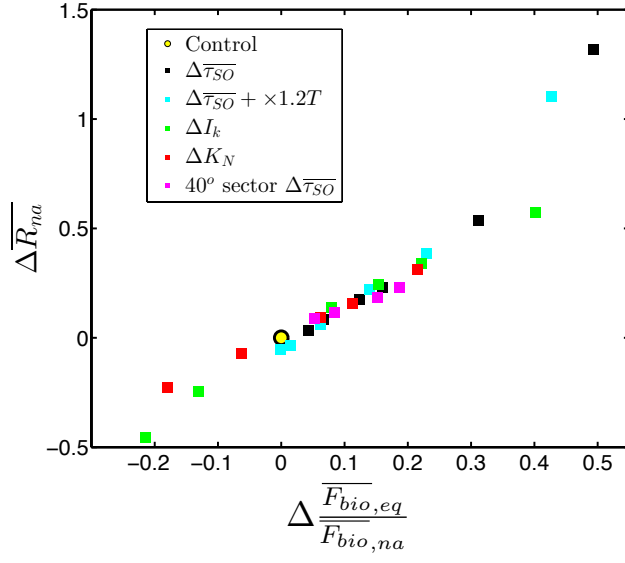
The results are summarised in Fig. 4.8. Firstly, Fig. 4.8a shows that $\overline{R_{na}}$ increases with $\frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$ and is linear for changes in $\frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$ less than 0.3. Within the linear regime, all

experiments follow the linear relationship stated above in eqn 4.7. (For the wind stress and radiative forcing experiments, atmospheric $p\text{CO}_2$ and buffer factor values from the the equilibrated $1.2 \times T$ experiment are taken as the control mean.)

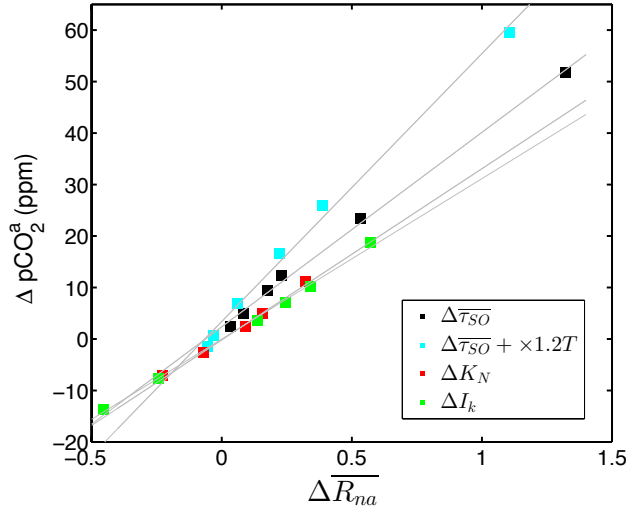
Secondly, Fig. 4.8b shows that $\Delta p\text{CO}_2^o$ varies linearly with $\Delta \overline{R_{na}}$ as predicted by eqn. 4.7. The MITgcm experiments therefore show that that an increase in Tropical biological activity relative to Northern sub-polar gyre biological activity results in an increase in atmospheric $p\text{CO}_2$, under a variety of forcing scenarios. However, the sensitivity of the response varies in each set of experiments. We can explain the variation by rearranging the definition of the buffer factor R (eqn. 4.2):

$$\delta p\text{CO}_2^o = R \frac{p\text{CO}_2^o}{\text{DIC}} \delta \text{DIC}. \quad (4.8)$$

The sensitivity of atmospheric $p\text{CO}_2$ to the Buffer Factor R will therefore depend on the background ratio $p\text{CO}_2^o/\text{DIC}$. The increase in sensitivity of atmospheric $p\text{CO}_2$ to the North Atlantic sub-polar Buffer Factor between the nutrient-limitation and wind stress experiment is due to the increased outgassing in the Southern Ocean. In the warming and wind stress experiments, the warming elevates the background $\overline{p\text{CO}_2^o}$ due to the reduced solubility while simultaneously reducing the local DIC due to dynamical effects (see chapter 5). The ratio $p\text{CO}_2^o/\text{DIC}$ is therefore increased giving a stronger sensitivity. The pre-conditioning of the North Atlantic through the ratio $p\text{CO}_2^o/\text{DIC}$ is thus setting the sensitivity of $p\text{CO}_2$ to the Buffer Factor.



(a)



(b)

Figure 4.8: a) Change in $\overline{R}_{na}$ as function of $\frac{\overline{F}_{bio,eq}}{\overline{F}_{bio,na}}$ as calculated from wind stress alone over the range $1-3 \times \tau_{SO}$ (black squares), wind stress alone in the 40° longitude model (magenta), wind stress and temperature (cyan squares), light half-saturation constant (green squares) and nutrient half-saturation constant (red squares) perturbation experiments with the MITgcm. b) Change in pCO_2^a as function of $\Delta \overline{R}_{na}$ for wind stress alone (black squares), wind stress and temperature (cyan squares), nutrient half-saturation constant (red squares) and light half-saturation constant (green squares) perturbations experiments with MITgcm.

The adjustment timescale of $\Delta \overline{pCO_2^a}$ in the North Atlantic shown in Fig. 4.1 is limited by the accumulation of biological material in the equatorial shadow zone, which takes

multiple centuries to equilibrate, similarly to the adjustment time of the $\overline{R_{na}}$ and ΔpCO_2^a . However, the residual circulation in the Southern Ocean equilibrates within two centuries so the Southern Ocean pCO_2 is not solely responsible for the pCO_2 anomaly.

To quantify the strength of the non-local climate feedback of atmospheric pCO_2 due to Southern Ocean wind stress changes, we need to quantify the effect of the North Atlantic sub-polar gyre pCO_2 change on atmospheric pCO_2 , due to a change in Southern Ocean winds. To put the atmospheric pCO_2 feedback of the sub-polar gyre into perspective, we compare it with the local Southern Ocean pCO_2^a feedback using the feedback parameter ϵ_f which follows the linear equations derived in section 2.4.2:

$$\Delta \overline{pCO_{2NA}^o} = \epsilon_f \Delta \overline{pCO_{2SO}^o}, \quad (4.9)$$

$$\Delta pCO_2^a = (1 + \epsilon_f \frac{P_{NA}}{P_{SO}}) P_{SO} \Delta \overline{pCO_{2SO}^o} + \Delta P_{SO} \overline{pCO_{2SO}^o} + \Delta P_{NA} \overline{pCO_{2NA}^o} + \Delta P_E \overline{pCO_{2E}^o}, \quad (4.10)$$

where the subscript SO refers to the Southern Ocean where the perturbation is applied, the subscript NA refers to the North Atlantic sub-polar gyre and the subscript E refers to the rest of the ocean basin⁹ where no dynamical change in pCO_2 is observed as a result of Southern Ocean wind perturbations (see below). P_{SO} , P_{NA} and P_E are the area fractions of the respective regions, weighed by the local wind stress and Δ indicates the anomaly with respect to the control mean. We will estimate ϵ_f using both eqns. 4.9 and 4.10. The values of P_{SO} , P_{NA} , ΔP_{SO} , ΔP_E and ΔP_{NA} can be explicitly calculated, while the values of ΔpCO_2^a , $\Delta \overline{pCO_{2SO}^o}$, $\Delta \overline{pCO_{2NA}^o}$, $\overline{pCO_{2SO}^o}$ and $\overline{pCO_{2NA}^o}$ can be diagnosed from the MITgcm control and wind stress perturbation experiments.

The results of the calculations are shown in Fig. 4.9. To illustrate the significance of the non-local feedback, Fig. 4.9a shows the value of the full atmospheric pCO_2 and the value of atmospheric pCO_2 calculated using the scaling from section 3.1.2 (i.e. the local feedback), $(pCO_2^a)_{SO}$, as a function of wind stress. In all cases, the full change in atmospheric pCO_2 is between 1.5 to 2 times larger than the change in $(pCO_2^a)_{SO}$, illustrating the significance of the non-local North Atlantic feedback.

Figure 4.9b shows the values of ϵ_f as function of Southern Ocean wind stress, as calculated by eqns. 4.9 and 4.10 after 400 years. While there are differences in the results

⁹The area 40°S to 40°N

between the two methods, the order of magnitude and trend of the calculated values of ϵ_f are similar. In the appendix (section 7.8), we outline a more rigorous method of estimating ϵ_f , which gives a similar trend but with higher overall values of ϵ_f . Values of ϵ_f for Southern Ocean wind stress are positive and between 0.5–0.8 for small perturbations in wind stress, and increase as the wind stress increases. The climate of feedback of Southern Ocean wind stress on atmospheric $p\text{CO}_2$ is therefore almost doubled by the non-local effect in $\overline{R_{na}}$.

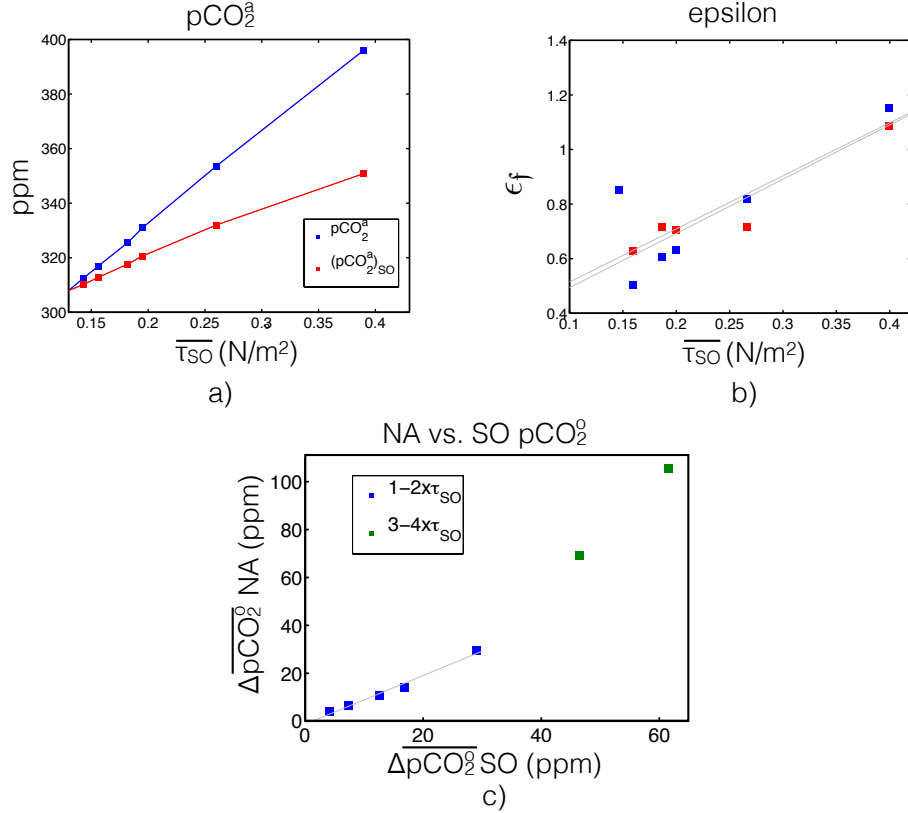


Figure 4.9: a) $p\text{CO}_2^a$ after 400 years as function of Southern Ocean wind stress due to Southern Ocean local effects alone (red) and including the non-local contribution (blue). b) ϵ_f as a function of wind stress in MITgcm estimated using eqns. 4.9 (blue) and 4.10 (red). The grey lines show the linear trend. c) North Atlantic $\Delta p\text{CO}_2^a$ against Southern Ocean $\Delta p\text{CO}_2^a$ for different Southern Ocean wind stress perturbation experiments in MITgcm.

In section 2.4.2, we showed that the sign of ϵ_f depends on the relative magnitude of the biogeochemical feedback described in this chapter (a positive feedback) and the redistribution of carbon within the system (as carbon concentrations in Southern Ocean go up, carbon concentrations elsewhere must go down through conservation of carbon: a

negative feedback), following eqn. 2.40:

$$\Delta \overline{pCO_2^o}_{2NA} = - \left(\phi + \overline{C_{SO}} \frac{\Delta \phi}{\Delta \overline{C_{SO}}} \right) \overline{R_{SO}} \frac{\overline{pCO_2^o}_{2NA}}{\overline{C_{NA}}} \left(\overline{R_{NA}} \frac{\overline{pCO_2^o}_{2SO}}{\overline{C_{SO}}} \right)^{-1} \Delta \overline{pCO_2^o}_{2SO}, \quad (4.11)$$

where $\overline{C_{SO}}$ and $\overline{C_{NA}}$ are the surface DIC concentrations in the Southern Ocean and North Atlantic respectively, and ϕ is a function of the strength of the MOC and vertical mixing (see eqn. 2.35). The first term in parentheses on the right hand side represents the re-distribution of carbon, whereas the second term measures the biogeochemical feedback. For the lowest wind stress perturbations, $\Delta \overline{\tau_{SO}} = 0.01 \text{ N/m}^2$, the biogeochemical feedback is only 10% stronger than the re-distribution feedback, so they mostly compensate each other. However, the relative magnitude of the biogeochemical feedback increases such that it is 55% times stronger than the re-distribution feedback at $\Delta \overline{\tau_{SO}} = 0.55 \text{ N/m}^2$ (estimated by calculating the terms in eqn. 2.36). As a result, the value of ϵ_f increases with wind stress.

The approach does not account for effects from the mid-latitudes and Tropics. To address this concern, additional wind stress perturbation experiments are run with pCO_2^g fixed at the control value, to examine the pattern of pCO_2^g in the basin without any atmospheric trends. The result is that between 30°S and 40°N, there is a negligible change in pCO_2^g : *DIC* increases as more carbon is upwelled while an increase in biology reduces *DIC*. In this case, the two contributions mostly balance since there no change in pCO_2^g so that the sub-polar gyre is the dominant contributor to atmospheric pCO_2 perturbations outside of the Southern Ocean.

It is useful to revisit one of the assumptions made in section 2.4.2 when defining ϵ_f , namely that the North Atlantic $\Delta \overline{pCO_2^g}$ is proportional to the $\Delta \overline{pCO_2^g}$ of the Southern Ocean. Figure 4.9c therefore shows North Atlantic $\Delta \overline{pCO_2^g}$ as a function of Southern Ocean $\Delta \overline{pCO_2^g}$ in the wind stress perturbation experiments. Over a range of perturbations up to $2 \times \tau_{SO}$ corresponding to a $\Delta \overline{\tau_{SO}}$ of 0.13 N/m^2 , the relationship is linear. The models of the CMIP5 archive all predict a $\Delta \overline{\tau_{SO}}$ of less than 0.13 N/m^2 for $2 \times CO_2$ experiments, so the assumption we made is valid for the foreseeable future climate.

We can apply eqn. 4.10 and our estimates of ϵ_f from the MITgcm experiments to predict the response of atmospheric pCO_2^g in the fully coupled global models from the

CMIP5 archive. We therefore repeat the calculation for Fig. 3.9 using a non-zero ϵ_f . The result of the calculation is shown in Fig. 4.10, which also includes the calculations for $\epsilon_f = 0$ from Fig. 3.9 for comparison. Figure 4.10 shows the effect of the non-local feedback mechanism induced by the wind-stress on atmospheric $p\text{CO}_2$ in the MITgcm, along with estimates for the HadGEM2-ES and IPSL-CM5. The non-local feedback increases the total climate equilibrium feedback of $\Delta\overline{\tau_{SO}}$ by 100%. The additional $\Delta p\text{CO}_2^a$ that results from $\Delta\overline{R_{na}}$ therefore constitutes a significant fraction of the feedback which cannot be measured by considering changes in the Southern Ocean carbon sink alone.

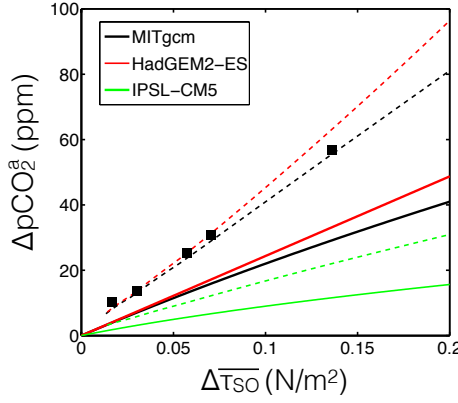


Figure 4.10: The change in atmospheric $p\text{CO}_2$ as estimated from the scaling in eqn. 4.10 using $\epsilon_f = 0$ as before in Fig. 3.9 (solid lines), and using the values of ϵ_f as estimated from the MITgcm experiments (dashed lines) for MITgcm (black) HadGEM2-ES (red) and IPSL-CM5 (green). The black squares are the MITgcm model output.

In this study we focus on the response of the North Atlantic buffer factor $\Delta\overline{R_{na}}$ to Southern Ocean wind stress perturbations $\Delta\overline{\tau_{SO}}$. The other side of the coin is true as well: if there is a decrease in equatorial biology, $\Delta\overline{R_{na}}$ will be negative and atmospheric $p\text{CO}_2$ will decrease. The nutrient and light limitation experiments in Fig. 4.8b and 4.8c illustrate the decrease in $\Delta\overline{R_{na}}$ and atmospheric $p\text{CO}_2$ as the ratio $\overline{F_{bio,eq}}/\overline{F_{bio,na}}$ decreases. In the real ocean, changes in stratification, wind-driven mixing, nutrient supply from elsewhere, local ecology [Weber and Deutsch, 2010] and even carbon concentrations [Riebesell et al., 2007] can affect the export of biological carbon. The effect of some factors such as temperature are less clear [Laws et al., 2010], but it is thought that increases in temperature decrease export of biological carbon [Segschneider and Bendtsen, 2013]. A global scale prediction of biological activity is therefore extremely challenging [Passow and Carlson, 2012], but our results show that the patterns of how biological productivity responds to

climate change can have significant impacts on atmospheric $p\text{CO}_2$. As Fig. 4.8c shows, changes in $\Delta\overline{R_{na}}$ of only 0.5 can cause a change in $p\text{CO}_2^a$ of 10 – 20 ppm.

While the analysis was performed using the MITgcm idealised set-up, we will attempt to identify a similar process in fully coupled global models. Lauderdale [2010] performs Southern Ocean wind stress perturbation experiments, similar to the ones presented here, with an MITgcm set-up with realistic bathymetry and more realistic temperature and salinity restoring and wind stress profiles. While not addressed in Lauderdale [2010], the experiments show a reduction in the North Atlantic sub-polar gyre carbon sink on long timescales of the same order of magnitude as the reduction in the Southern Ocean, shown in Fig. 4.11, possibly supporting our proposed mechanism.

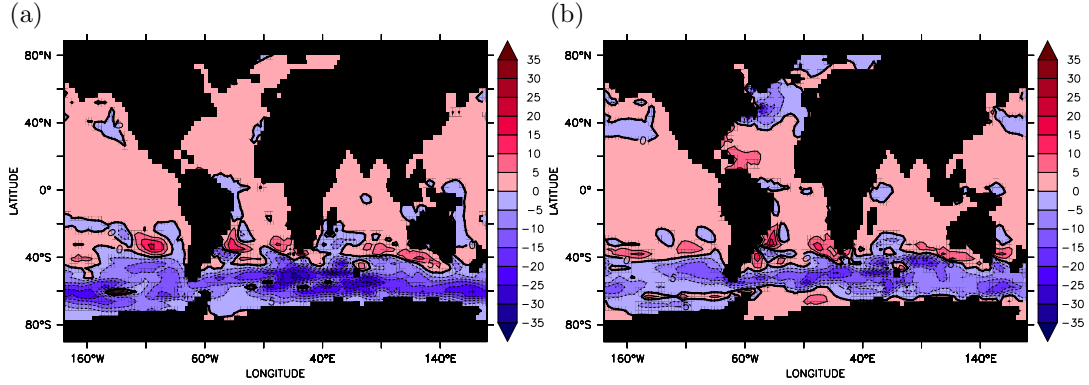


Figure 4.11: DIC tendency anomaly ($\text{mmol C m}^{-3} \text{ yr}^{-1}$) due to the air-sea carbon flux for a 50% increase in Southern Ocean wind stress after a) 10 years and b) 5000 years. Negative anomalies represent a reduction in ocean uptake of carbon, from Lauderdale [2010].

In fully coupled climate models such as HadGEM2-ES, we use the abrupt $4\times\text{CO}_2$ experiment,¹⁰ which is only 140 years long. The integration to 140 years is much less than the equilibration timescale of the North Atlantic buffer factor adjustment, but we can still observe the transient signal. We choose to focus on the first 100 years of the experiment, since the changes in the ocean system after 100 years become relatively smaller but are nonetheless present. HadGEM2-ES has a nutrient stream in the Atlantic Ocean with the maximum transport occurring along the $\sigma = 26 \text{ kg/m}^3$ density surface. The control mean nutrient stream S_P is shown in Fig. 4.12a, as well as the depth of the density surface $\sigma = 26 \text{ kg/m}^3$ in the control run (Fig. 4.12b) and after 100 years for the $4\times\text{CO}_2$ experiment

¹⁰In the abrupt $2\times\text{CO}_2$, there are no large scale patterns in the change in biological productivity, nor a change in the ratio $\frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$ therefore we only present the results of the $4\times\text{CO}_2$ experiment.

(Fig. 4.12c).¹¹

In HadGEM2-ES, the control run climatological ratio of equatorial biological productivity $\overline{F_{bio,eq}}$ over North Atlantic biological productivity, $\overline{F_{bio,na}}$, $\frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$, is roughly 2.4. The ratio $\frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$ is higher than estimated from observational chlorophyll data, which is less than 1 (using NODC chlorophyll data). In the $4\times\text{CO}_2$ experiment, both $\overline{F_{bio,eq}}$ and $\overline{F_{bio,na}}$ weaken (see Fig. 4.12d) but not by the same amount so that $\Delta \frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$ is roughly +1.1 (measured as the difference between a 10-year average centred around year 100 of the $4\times\text{CO}_2$ experiment and the control run climatology). Due to the different magnitudes in the response of $\overline{F_{bio,eq}}$ and $\overline{F_{bio,na}}$ and other dynamical and biogeochemical changes in the model, a quantitative analysis will be limited and we only examine patterns in a qualitative sense. In the HadGEM2-ES $4\times\text{CO}_2$ experiment, the clearest signal is observed in the Alkalinity changes,¹² which we will therefore focus our attention on.

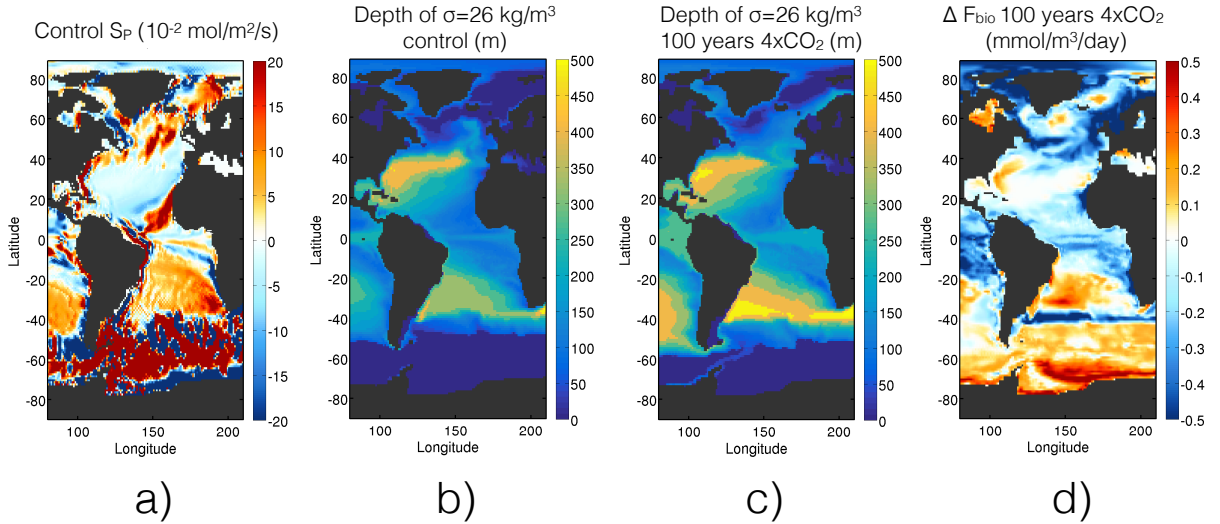


Figure 4.12: HadGEM2-ES: a) Nutrient stream S_p calculated from the control run climatology along the $\sigma = 26 \text{ kg/m}^3$ density surface, b) depth of the $\sigma = 26 \text{ kg/m}^3$ density surface in the control experiment, c) depth of the $\sigma = 26 \text{ kg/m}^3$ density surface after 100 years of the $4\times\text{CO}_2$ experiment and d) change in depth-averaged mixed layer biological productivity after 100 years of the $4\times\text{CO}_2$ experiment.

As discussed in section 4.1.2, a local increase in surface biology leads to a local decrease in Alkalinity below the thermocline. Conversely, a reduction in surface biology, such as in

¹¹The nutrient stream in the $4\times\text{CO}_2$ experiment is not shown because the nutrient field for the $4\times\text{CO}_2$ experiment was not available.

¹²The modelled changes in the Buffer Factor R in the HadGEM2-ES $4\times\text{CO}_2$ experiment are very noisy at high latitudes are strongly affected by anthropogenic CO_2 anomalies.

the equatorial shadow region in the HadGEM2-ES $4\times\text{CO}_2$ experiment, leads to an increase in Alkalinity below the thermocline. The change is most apparent in the regenerated Alkalinity, ALK_{reg} , which results from the biological flux at depth¹³ (similar in concept to regenerated carbon). Figures 4.13a-d therefore show the change in regenerated Alkalinity along the $\sigma = 26 \text{ kg/m}^3$ density surface after 5, 20, 50 and 100 years of the $4\times\text{CO}_2$ experiment. After 5 years, no obvious pattern is apparent. However, after 20 years, we see an increase in ALK_{reg} in the equatorial shadow zone (region 1 in Fig. 4.13) where there is a modelled decrease in the surface biological productivity, consistent with our hypothesis. We also observe that the positive ALK_{reg} anomaly from region 1 is advected along the nutrient stream S_P in the western side of the Northern hemisphere (region 2 in Fig. 4.13). The positive ALK_{reg} anomalies in both regions 1 and 2 strengthen with time. The ALK_{reg} anomalies continue to increase after 100 years, consistent with centennial timescale of the mechanism in MITgcm. Overall, the pattern in the ALK_{reg} anomaly is very similar to the pattern in the Buffer Factor anomalies shown in Fig. 4.5.

As the positive ALK_{reg} anomaly is advected along the western boundary current in the Northern sub-tropical region, we see that the signal disappears in the extension of the Gulf Stream. At this point, the $\sigma = 26 \text{ kg/m}^3$ density surface shoals towards the surface and the ALK_{reg} anomaly is mixed with surface waters. Due to definition of regenerated and preformed quantities, the regenerated Alkalinity is re-labeled preformed Alkalinity ALK_{pre} . Figure 4.13e-h therefore shows the change in preformed Alkalinity along the $\sigma = 26 \text{ kg/m}^3$ density surface after 5, 20, 50 and 100 years of the $4\times\text{CO}_2$ experiment. For each snapshot, the ALK_{reg} anomaly disappears where a positive ALK_{pre} anomaly appears in the extension of the Gulf Stream (see region 3 in Fig. 4.13). Subsequently, the positive ALK_{pre} anomaly is advected both into the sub-polar region where it is partially eroded by the local decrease in biology, and back into the sub-tropical gyre region.

To illustrate the effects of the Alkalinity changes on the sub-polar gyre $p\text{CO}_2$ and air-sea carbon flux, Fig. 4.14 shows the anomalies in ALK_{pre} on the $\sigma = 26 \text{ kg/m}^3$ density surface (panel a), surface ocean $p\text{CO}_2$ (panel b) and the air-sea carbon flux (panel c) between year 50 and 20. The difference between year 50 and 20 is shown for all quantities because the large atmospheric CO_2 step perturbation means patterns are obscured if we

¹³Because surface biology reduces Alkalinity below the thermocline, ALK_{reg} at depth can be negative. We calculate ALK_{reg} and ALK_{pre} using a regression onto salinity which may introduce errors at depth [Broecker, 1974].

use the control mean as a reference for ocean $p\text{CO}_2^o$ and the air-sea flux. Fig. 4.14 shows that areas with a positive ALK_{pre} anomaly correspond to negative anomalies in $p\text{CO}_2^o$. The net air-sea carbon flux over the North Atlantic decreases between year 20 and 50, since the yearly atmospheric input of CO_2 decreases over time. Negative anomalies in $p\text{CO}_2^o$ in Fig. 4.14b therefore correspond to smaller reductions in the local air-sea carbon flux. Similarly, negative ALK_{pre} anomalies correspond to strong reductions in the air-sea carbon flux. The patterns are most apparent in the areas marked by arrows in panel in Fig. 4.14.

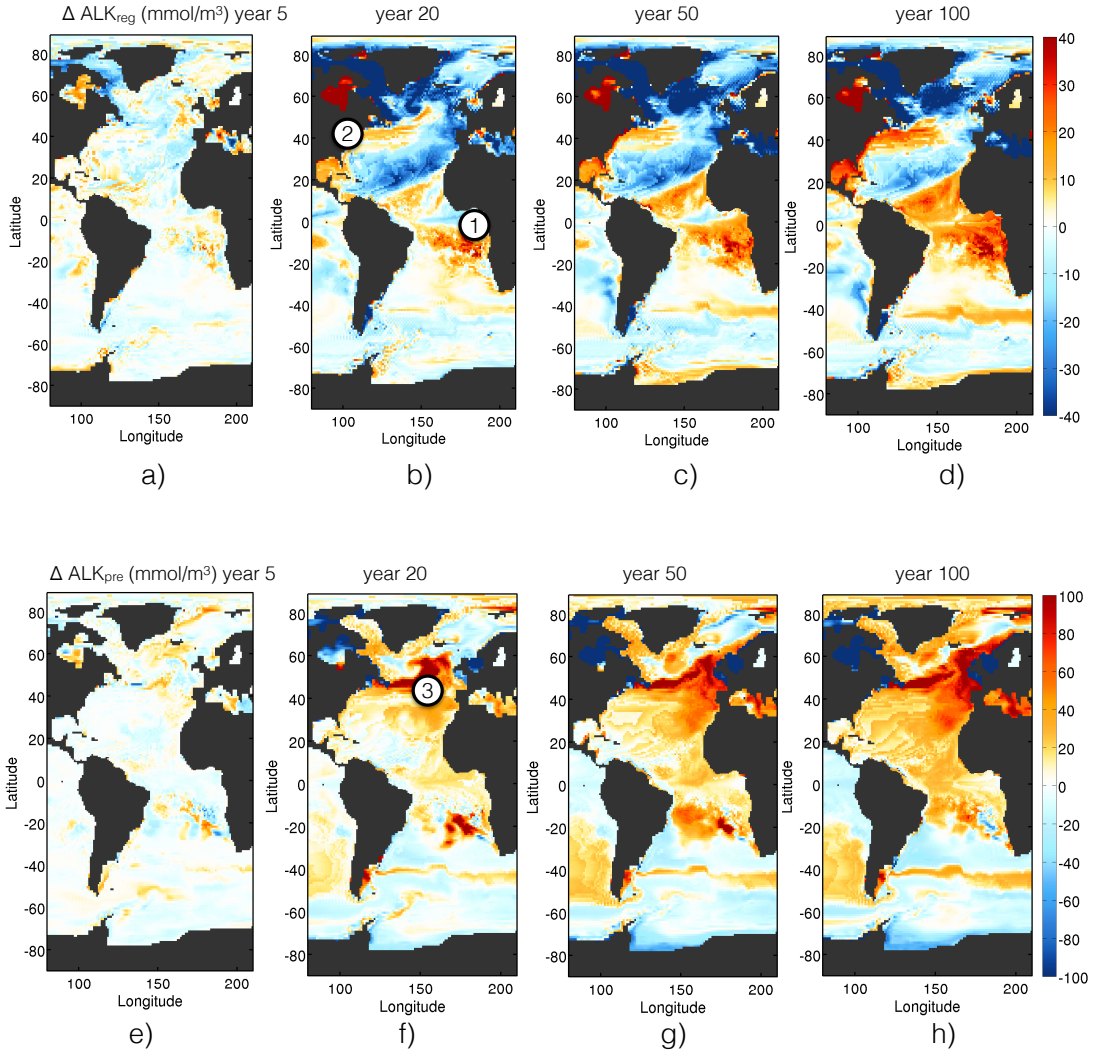


Figure 4.13: HadGEM2-ES: change in regenerated Alkalinity along the $\sigma = 26 \text{ kg/m}^3$ density surface after a) 5, b) 20, c) 50 and d) 100 years of the $4\times\text{CO}_2$ experiment relative to the control. Change in preformed Alkalinity along the $\sigma = 26 \text{ kg/m}^3$ density surface after e) 5, f) 20, g) 50 and h) 100 years of the $4\times\text{CO}_2$ experiment.

While the observed patterns in ALK_{reg} , ALK_{pre} , pCO_2 and $F_{air-sea}$ HadGEM2-ES do not verify the relationship between equatorial biology and sub-polar gyre Buffer Factor shown in MITgcm, the patterns are consistent with the mechanism discussed in this chapter: a change in equatorial biology causes changes in the chemistry of the sub-thermocline waters in the shadow zone, which are subsequently advected into the nutrient stream and into the Northern sub-polar gyre. The changes in Alkalinity and DIC advected into the sub-polar gyre from the equatorial shadow zone then affect local chemistry and CO_2 uptake. The positive ALK_{pre} anomaly (region 3 in Fig. 4.13) is advected across a large region, so the chemical changes potential affect a large area and could have consequences beyond ocean CO_2 uptake alone (such as local ocean acidification).

The work presented shows the influence of equatorial biology on the North Atlantic carbon sink and quantifies the atmospheric climate-carbon feedback of the equatorial biology due to the Southern Ocean winds. To put the result into perspective, we can quantify the atmospheric climate feedback of North Atlantic sub-polar winds. The *non-local* atmospheric feedback of the Southern Ocean is therefore compared to the *local* atmospheric feedback of the North Atlantic sub-polar winds.

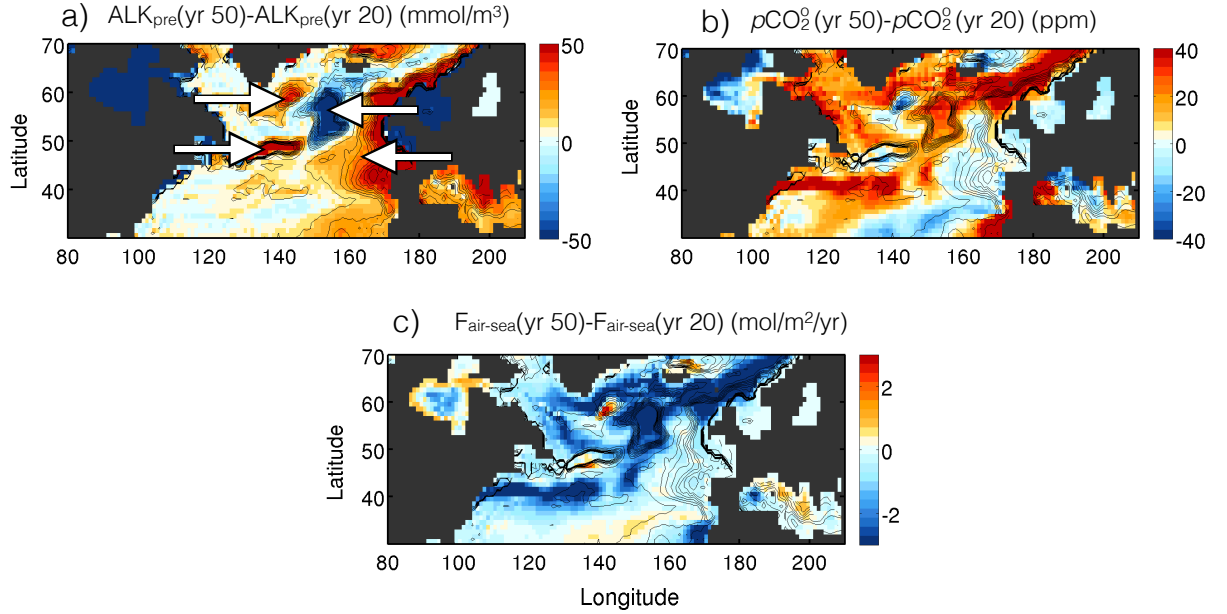


Figure 4.14: HadGEM2-ES anomalies of a) ALK_{pre} on the $\sigma = 26\ kg/m^3$ density surface, b) surface ocean pCO_2 and c) the air-sea carbon flux $F_{air-sea}$ between year 50 and 20 of the $4\times CO_2$ experiment. The black lines show the contours of the ALK_{pre} anomaly in panel a. The anomalies are taken using 5-year means around year 50 and 20.

4.2 Climate impact of the local North Atlantic sub-polar gyre wind

Similarly to the Southern Annular Mode (SAM), the winter-time Northern Annular Mode (NAM) is projected to increase under climate change in CMIP5 simulations [Karpechko, 2010, Yali and Huijun, 2010, Gillett and Fyfe, 2013], although the spread in projections is large and model-dependent [Karpechko, 2010]. The increase in the NAM index is expected to be only roughly half that of the SAM by 2100 [Gillett and Fyfe, 2013]. Much like the SAM, a stronger positive NAM is coupled to stronger zonal winds due to an increase in the meridional atmospheric pressure gradient.

However, little work has been done on the impact of increased zonal winds in the Northern high-latitudes on ocean carbon uptake and subsequent atmospheric CO₂ levels. Much like the Southern Ocean, the North Atlantic sub-polar gyre sees deep wintertime mixed layer depth and convective mixing, providing a window to the deep ocean. A wind-induced perturbation in vertical mixing could therefore expose deep carbon to the atmosphere. It was shown in section 4.1 that the atmospheric carbon feedback of zonal winds over the Southern Ocean can be substantial through both local and non-local effects. In this section we quantify the relative magnitude of the atmospheric carbon feedback due to zonal winds over the Atlantic sub-polar gyre and estimate the magnitude of the climate-carbon feedback. To that end, we employ the MITgcm and analyse the response of the system to perturbations in the zonal wind stress acting over the North Atlantic (between 20°-60°N). We will examine of the carbon cycle response to a range of wind stress perturbations, while specific examples will be drawn from the $2 \times \tau_{NA}$ scenario.

4.2.1 Model response to increased zonal wind stress

An increase in area-mean North Atlantic zonal wind stress¹⁴ $\overline{\tau_{NA}}$ causes a strengthening of the Northern sub-tropical and sub-polar gyres due to increased input of wind stress curl (Fig. 4.15a, Fig. 4.16) [Munk, 1950]. The increase in gyre circulation results in an increase in northward heat transport. There is increased mixing associated with the stronger winds deepening the mixed layer. However, the elevated temperatures in the sub-polar gyre (particularly in the extension of the western boundary current and North-Eastern corner, see Fig. 4.16) increase stratification and reduce the overall mixed layer

¹⁴Averaged over the area 0°-20°E and 40°-60°N.

depth (Fig. 4.15c) [Lohmann et al., 2009]. The increase in northward heat transport also results in an increased in ocean heat uptake in the North Atlantic since deep waters formed at the Northern high latitudes are warmer. The buoyancy of the upper ocean is therefore increased (between 0-1500m depth) leading to a deepening of pycnocline as the strength of the North Atlantic wind stress increases.

The increase in northward heat transport and subsequent increase in heat loss of northward advected water parcels also results in an increase in convection and a weak strengthening of the MOC [Eden and Willebrand, 2001, Lohmann et al., 2009]. However, the strengthening in the MOC is concentrated to in the Northern hemisphere only and does not extend to the Southern channel (Fig. 4.16a). The increase in upper ocean heat uptake increases the vertical density gradient across the basin, including the Southern Ocean, causing a slight reduction in the strength of the bottom overturning cell (Fig. 4.15b).

The changes in both the meridional overturning circulation and the barotropic circulation are manifested in the tracer fields. Figure 4.17a-b show the anomalies in potential temperature induced in the $2 \times \tau_{NA}$ experiment. Increased Ekman convergence in the subtropical gyre results in an increase in downwelling and uptake of heat while the increased Ekman divergence in the sub-polar gyre upwells cold waters. The increase in gyre circulation increases northward heat transport, resulting in warming of the Eastern sub-polar gyre, while a stronger western boundary current advects more cold water South and cools the Eastern edge. The combination of lateral and vertical heat transport changes lead to a dipole pattern in the zonal mean temperature response between 0-700m depth. As mentioned previously, the increase in northward heat transport also results in increase ocean heat uptake in the upper ocean between 0-1500m depth.

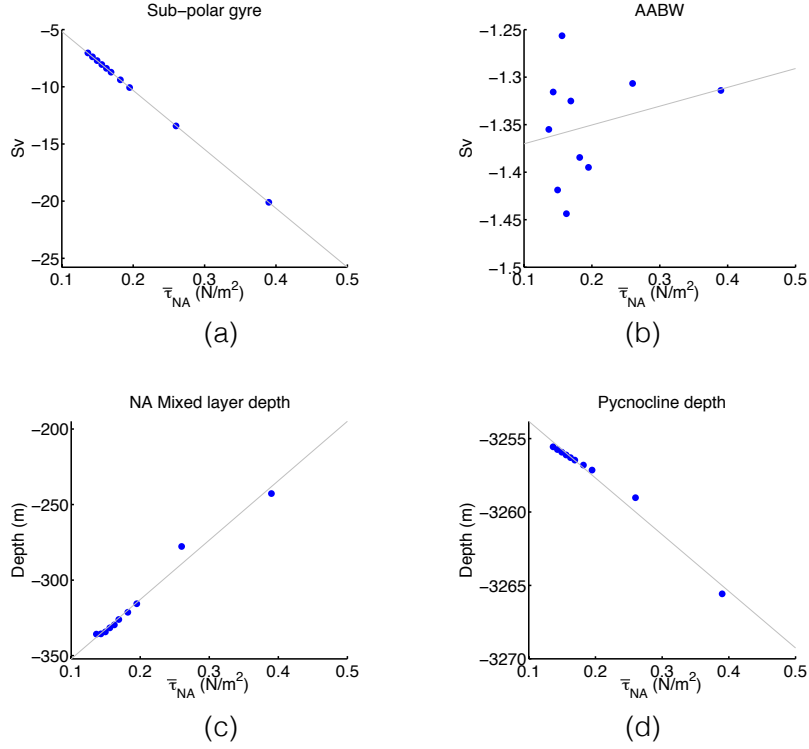


Figure 4.15: MITgcm diagnostics as a function of area-mean North Atlantic sub-polar gyre wind stress (40°-60°N): a) Minimum in sub-polar gyre barotropic stream function (Sv), b) Minimum in the bottom overturning cell, AABW (Sv), c) area-mean mixed layer depth (m, 40°-60°N) and d) pycnocline depth (m).

The dissolved inorganic carbon distribution generally follows the temperature distribution at depth (Fig. 4.17c-d). The increase in sub-tropical gyre downwelling and northwards lateral transport reduces underlying DIC concentrations due to mixing of water parcels with low DIC concentrations. The increase in upwelling and mixing in the sub-tropical gyre brings carbon-rich waters to the surface. The increase in ocean heat uptake causes an increase in stratification throughout the basin isolates carbon below 1500m depth.

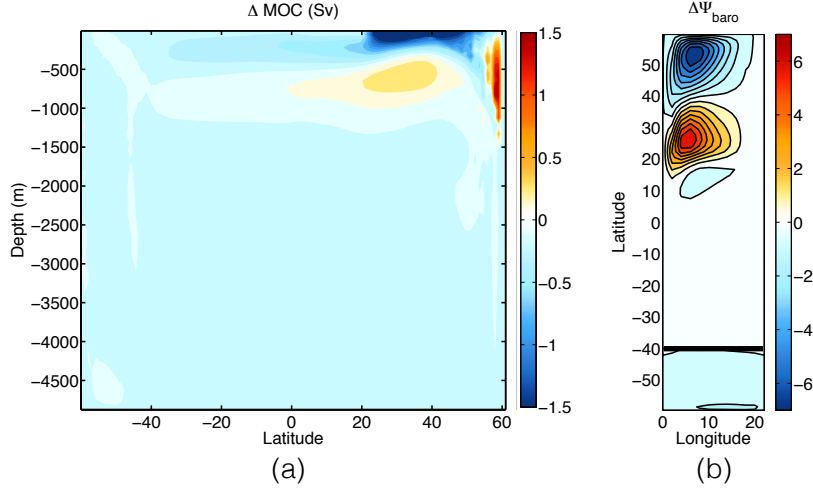


Figure 4.16: Change in a) meridional overturning streamfunction and b) barotropic streamfunction after 400 years in the $2 \times \tau_{NA}$ experiment relative to the control mean (Sv). The thick black line in panel b indicate the extent of the Southern channel.

Due to the increased Ekman divergence in the North Atlantic sub-polar gyre and the increase in the Northern hemisphere MOC, more nutrients from the upper ocean are re-cycled within the Northern hemisphere. The southward transport of nutrients across the equator is therefore reduced (Fig. 4.17e). At the surface, we therefore see an increase in nutrients in the Northern hemisphere and a decrease in the Southern regions (Fig. 4.17f) causing an increase and decrease in biological productivity, respectively. Due to the adaptations of the surface biology, the changes in nutrients in the upper cell are reflected in the lower cell through the vertical rain of biological material (Fig. 4.17e).

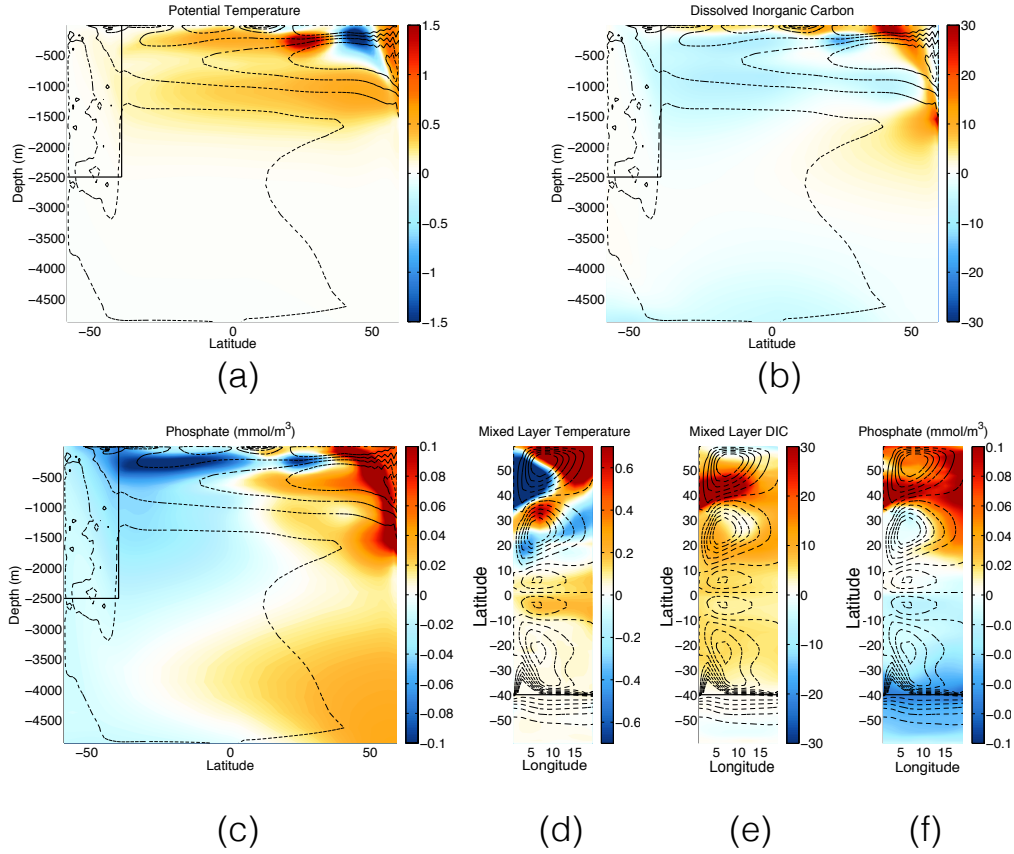


Figure 4.17: MITgcm output of zonal mean anomalies of a) potential temperature, b) dissolved inorganic carbon and c) phosphate, and mixed layer anomalies of d) potential temperature, e) dissolved inorganic carbon and f) phosphate in response to a $2 \times \tau_{NA}$ perturbation. All anomalies are taken as the difference between an average over the last 10 years of a 400 year-long run and a control mean. Dashed lines show the perturbed overturning stream function (panels a-c) and perturbed barotropic stream function (d-f). Solid black lines indicate the location of the Southern channel.

The net change in ocean carbon content is dictated by the release of carbon from the sub-polar gyre following the increase in wind-driven mixing. The result is an increase in atmospheric CO_2 as the North Atlantic wind stress increases (Fig. 4.18a). The atmospheric CO_2 anomalies as a result of a North Atlantic wind stress perturbation are 50% smaller than those associated with a Southern Ocean wind stress perturbation. The change is driven by adjustment of the North Atlantic sub-polar gyre $\overline{p\text{CO}_2}$ (Fig. 4.18b), which has an initial adjustment time of 10 years, after which the pycnocline deepening provides a second centennial timescale of adjustment. Finally, the trapping of regenerated carbon in the bottom cell causes a millennial timescale decrease in the sub-polar gyre $p\text{CO}_2$. Atmospheric $p\text{CO}_2$ follows the response of the sub-polar gyre (Fig. 4.18b).

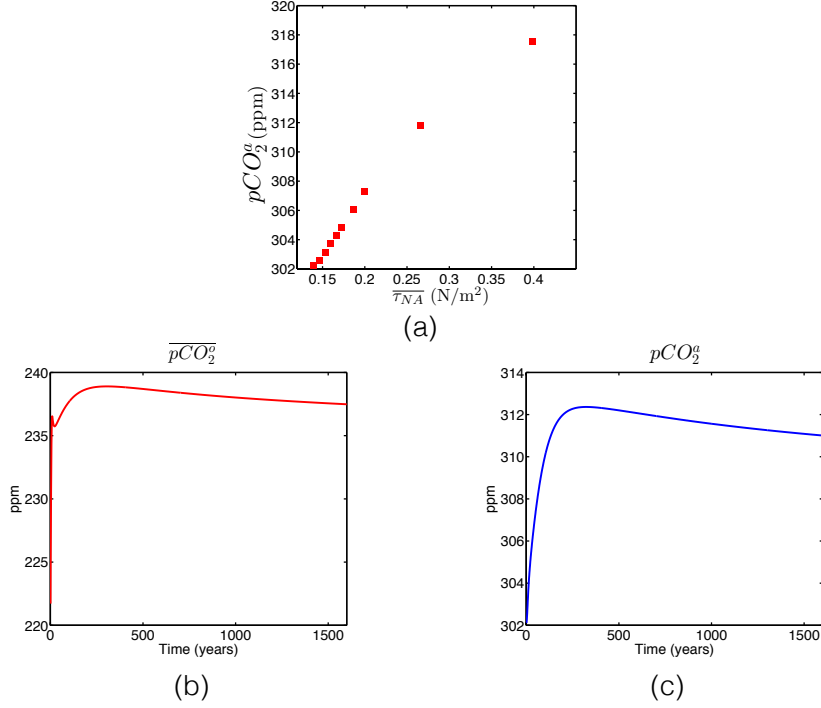


Figure 4.18: a) Atmospheric pCO_2 in MITgcm as a function of area mean North Atlantic sub-polar gyre wind stress $\overline{\tau_{NA}}$ after 1600 years of experiment, b) North Atlantic ocean pCO_2 (40°-60°N) and c) atmospheric pCO_2 as a function of time for the $2 \times \tau_{NA}$ experiment.

Similar to the Southern Ocean, we can apply the mixed layer carbon budget derived in section 2.4 to examine and predict the magnitude of local ocean and global atmospheric pCO_2 anomalies.

4.2.2 Evaluating the climate feedback of North Atlantic wind stress

To construct a relationship for the North Atlantic sub-polar gyre in terms of the wind stress in the North Atlantic sub-polar gyre τ_{NA} , we revisit the carbon mixed layer flux equation 2.26. Similar to section 3.1.2, we can express the air-sea flux $F_{air-sea}$ as $F_{air-sea} = K\tau_{NA}(pCO_2^a - pCO_2^o)$ and solve for pCO_2^o :

$$pCO_2^o = \left[\frac{K\tau_{NA}}{h} pCO_2^a - C_h(w_h + u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y}) - [F_{bio}]_h \right] / K\tau_{NA}. \quad (4.12)$$

Just as we did with the Southern Ocean mixed layer carbon equation, we express the mixed layer entrainment velocity as $w_h = -\frac{2}{\alpha g h \Delta_z T} \left(m_0 u_*^3 - \alpha g h \frac{Q}{2\rho_0 C_p} \right)$, with u_* as the frictional wind speed $u_* = \left| \frac{\tau_{NA}}{\rho_0} \right|^{1/2}$. Similarly, we approximate the residual meridional velocity as the Ekman velocity and a bolus velocity that is linearly proportional to the Ekman velocity via a parameter e_{GM} such that $v = (1 + e_{GM}) \frac{\tau_{NA}}{f\rho_0 h}$. The ocean pCO_2 can

then be written as:

$$pCO_2^o = - \frac{C_h}{K} \left(- \frac{2m_0}{\alpha gh \Delta_z T \rho_0^{3/2} \tau_{NA}^{1/2}} + \frac{Q}{\Delta_z T \rho_0 C_p \tau_{NA}} + \frac{(1 + e_{GM})}{\rho_0 f h} \frac{\partial h}{\partial y} + \frac{u}{\tau_{NA}} \frac{\partial h}{\partial x} \right) + \frac{[F_{bio}]_h}{K \tau_{NA}} + pCO_2^a. \quad (4.13)$$

We then apply a zonal and meridional weighted average over 0°-20°E and 40°-60°N, respectively (the average is denoted by an overbar in the rest of the section):

$$\begin{aligned} \overline{pCO_2^o} = & \overline{\left(\frac{C_h}{K} \right) \frac{2m_0}{\alpha gh \Delta_z T \rho_0^{3/2} \tau_{NA}^{1/2}}} - \overline{\left(\frac{C_h}{K} \right) \frac{Q}{\Delta_z T \rho_0 C_p \tau_{NA}}} \\ & - \overline{\left(\frac{C_h}{K} \right) \frac{(1 + e_{GM})}{\rho_0 f h} \frac{\partial h}{\partial y}} - \overline{\left(\frac{C_h}{K} \right) \frac{u}{\tau_{NA}} \frac{\partial h}{\partial x}} \\ & + \frac{\overline{[F_{bio}]_h}}{K \tau_{NA}} + pCO_2^a, \end{aligned} \quad (4.14)$$

Because the domain is not zonally symmetric like the Southern Ocean, we preserve the zonal advection term. In the North Atlantic case, we once again choose to calculate $\Delta_z T$ with respect to a depth of 1000m. Similarly to the Southern Ocean case, we label the first and second terms the (mechanical) mixing and the radiative flux (buoyancy mixing) terms, respectively, although they both represent a form of vertical mixing. Figure 4.19 shows the magnitude of the terms in eqn. 4.14 as a function of North Atlantic wind stress, as diagnosed from the MITgcm experiments. We find that the zonal advection and radiative flux terms are far more dominant in determining $\overline{pCO_2^o}$ in the North Atlantic than in the Southern Ocean, as a result of stronger zonal mixed layer gradients and heat fluxes in the North Atlantic. Figure 4.19f shows that $\overline{Q}$ becomes more negative with wind stress, which would increase $\overline{pCO_2^o}$ due to deepening of the mixed layer. The weakening of the radiative flux term with wind stress is therefore driven by the increase in τ_{NA} , which causes faster outgassing of ocean carbon. Similarly, the weakening of the zonal advection term is driven by the increase in τ_{NA} .

Apart from the vertical mixing term, which becomes more dominant with the strength of the wind stress, the remainder of the terms are negligible. The difference in the magnitude of the terms illustrates the different dynamical regimes in the Southern Ocean and North Atlantic sub-polar gyre. The vertical mixing term is initially diagnosed as negative due to small negative values of $\Delta_z T$ in the convecting grid cells in the North-Eastern edge of the domain. Similarly, these small values result in large changes in vertical mixing term

in the $2 \times \tau_{NA}$ experiment. To avoid these issues, we use the control mean values of $\Delta_z T$ away from the North-Eastern edge of the domain in our scaling to predict North Atlantic pCO_2^o . Figure 4.19 also shows the terms in eqn. 4.14 if we split the average to isolate $\overline{\tau_{NA}}$, like in section 3.1.2. For the North Atlantic case, we find that the procedure of splitting the integral and isolating $\overline{\tau_{NA}}$ introduces large errors in all terms. Therefore, we keep the terms as they appear in eqn. 4.14. Due to the dominance of the zonal advection and radiative flux terms, the scaling for $\overline{pCO_2^o}$ does not follow a simple square-root relationship. However, we can still use it to predict $\overline{pCO_2^o}$ as a function of $\overline{\tau_{NA}}$ if we include the vertical mixing, zonal advection and radiative flux terms. The result is the equation:

$$\begin{aligned} \overline{pCO_2^o} = & \overline{C_h \frac{2m_0}{K \alpha g h \cdot \Delta_z T \rho_0^{3/2} \tau_{NA}^{1/2}}} - \left(\frac{C_h}{K} \right) \frac{u}{\tau_{NA}} \frac{\partial h}{\partial x} \\ & - \frac{C_h}{K} \frac{Q}{\rho_0 \Delta_z T C_p \tau_{NA}} + pCO_2^a + \mathfrak{F}(u, v, F_{bio}), \end{aligned} \quad (4.15)$$

where the $\mathfrak{F}$ -term contains the terms whose variations with $\overline{\tau_{NA}}$ are small compared to the vertical mixing and radiative flux terms, such that:

$$\mathfrak{F}(v, F_{bio}) = - \left(\frac{C_h}{K} \right) \frac{(1 + e_{GM})}{\rho_0 f h} \frac{\partial h}{\partial y} + \frac{[F_{bio}]_h}{K \tau_{NA}}. \quad (4.16)$$

We diagnose K from the control simulation.¹⁵, giving a value of $K \overline{\tau_{NA}} = 7$ cm/hr which is within the range suggested by the literature [Wanninkhof, 1992].

To verify the relationship, Fig. 4.20a shows North Atlantic pCO_2^o as a function of area-mean wind stress, along with the calculation of $\overline{pCO_2^o}(\overline{\tau_{NA}})$ as given by eqn. 4.15, which matches the data well for small wind stress perturbations. The estimates from the MITgcm experiments and from the scaling for North Atlantic $\overline{pCO_2^o}$ do not follow a simple square root relationship and deviate from a linear trend due to the contribution of the vertical mixing, zonal advection and radiative flux terms. The magnitude of the $\overline{pCO_2^o}$ anomalies is also smaller in the North Atlantic than in the Southern Ocean. For example, the response to a 0.13 N/m^2 change in area mean wind stress gives a 17ppm change in $\overline{pCO_2^o}$ in the North Atlantic while the equivalent change in the Southern Ocean is 40ppm.

¹⁵ K is diagnosed by solving eqn. 4.15 for the control mean value of $\overline{pCO_2^o}$.

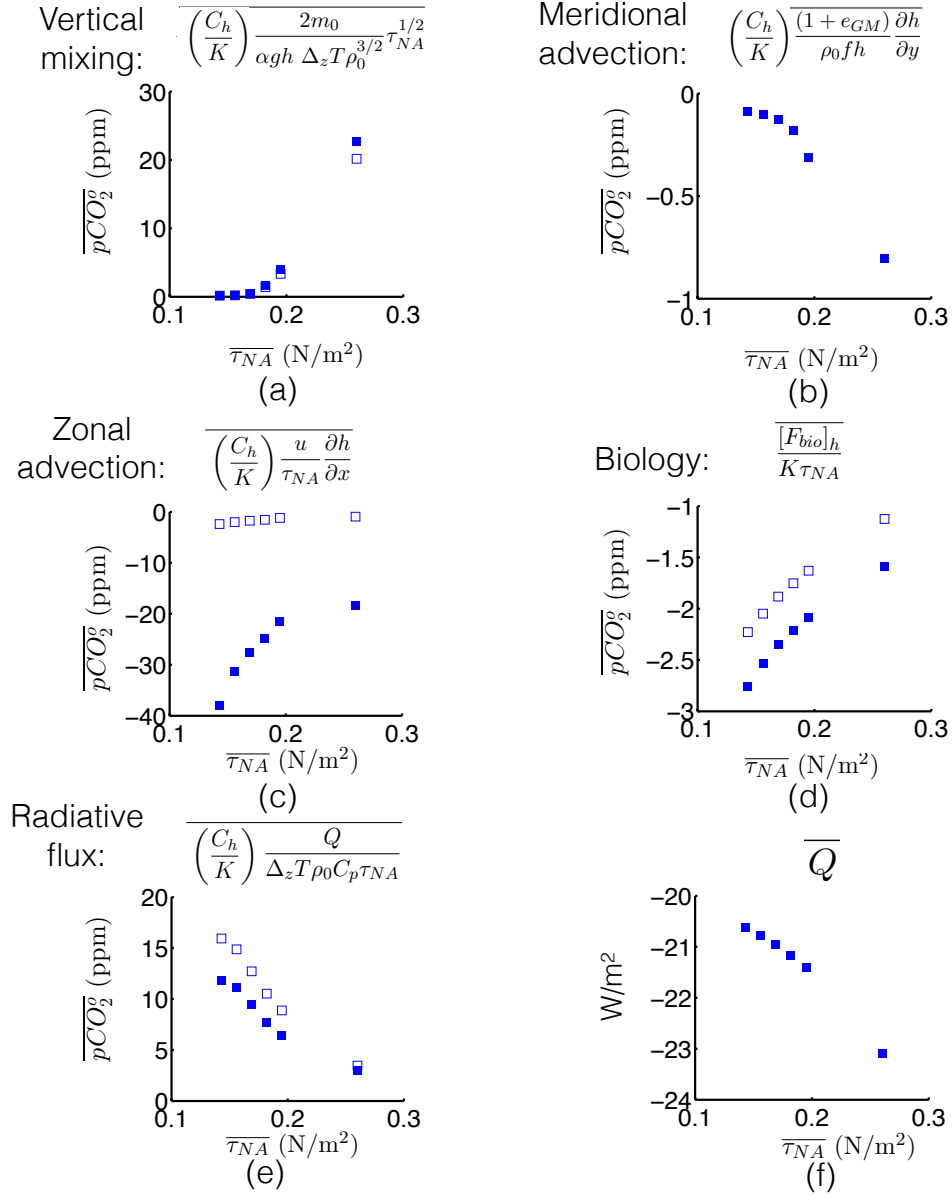


Figure 4.19: a)-e) Contribution of terms in the North Atlantic $p\text{CO}_2$ equation as a function of wind stress: a) vertical mixing, b) meridional advection (Ekman), c) zonal advection, d) biology and e) radiative flux contributions. The filled squares show the terms as they appear in eqn. 4.14 while the empty squares shows the terms with the average split to isolate $\overline{\tau_{NA}}$. Panel (f) shows the mean of the radiative flux Q with North Atlantic wind stress.

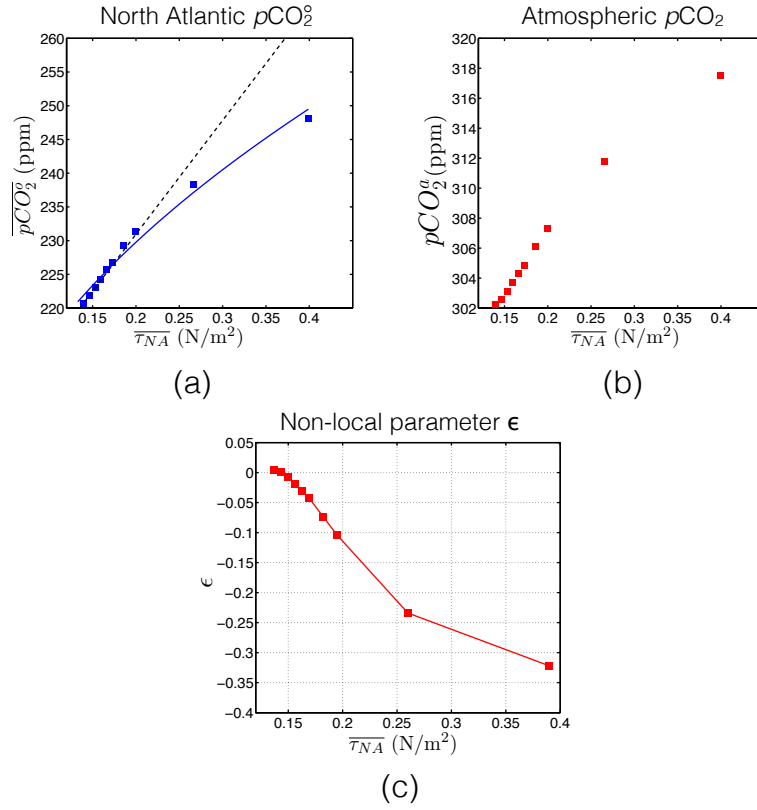


Figure 4.20: a) North Atlantic pCO_2^o in MITgcm as a function of wind stress after 1600 years of experiment (blue squares) with the calculation using the wind stress scaling based on control mean parameters (solid blue line) and a linear trend around the control state (dotted back line); b) Atmospheric pCO_2 in MITgcm as a function of area mean North Atlantic sub-polar gyre wind stress $\overline{\tau_{NA}}$ after 1600 years of experiment and c) the non-local feedback parameter ϵ_f a function of mean wind stress $\overline{\tau_{NA}}$ after 1600 years of experiment.

The difference in the overall sensitivity of the $\overline{pCO_2^o}$ to wind stress perturbations is a result of the magnitude of the parameters C_h , h and $\Delta_z T$ in the scaling equation 4.15. The values of these parameters give a reduced sensitivity to vertical mixing. The magnitude of the area-averaged parameters in the North Atlantic and the Southern Ocean are shown in table 4.2.2.

Variable	Southern Ocean	North Atlantic	Units
$\overline{h}$	130	357	m
$\overline{C_h}$	2.204	2.048	mol/m ³
$\overline{\Delta_z T}$	4.9	6.1	°C
$\overline{C_h \frac{1}{h \cdot \Delta_z T}}$	0.027	0.0026	mol/m ⁵ /°C

Figure 4.20b shows the atmospheric pCO_2 anomaly associated with the $\overline{\tau_{NA}}$ anomalies. In the Southern Ocean, there was as strong non-local feedback of the wind stress on atmospheric CO_2 levels due to the changes in the nutrient stream. Changes in the North

Atlantic wind stress do not show a similar strong non-local feedback however. We can use the same methodology outlined in section 4.2.2 to calculate the non-local feedback parameter ϵ_f using eqn. 4.10, outside the sub-polar gyre. The results of the calculation are shown in Fig. 4.20c. The magnitude of the non-local contribution is small, generally less than 20% for perturbations smaller than 0.13 N/m^2 but increasing with $\overline{\tau_{NA}}$, and generally negative.

In section 2.4.2, we showed that the sign of ϵ_f depends on the relative magnitude of positive or negative non-local feedbacks due to dynamical or biogeochemical changes and the re-distribution of carbon within the system: as carbon concentrations in Southern Ocean go up, carbon concentrations elsewhere must go down through conservation of carbon; a negative feedback. A negative dynamical feedback can arise if forcing perturbations applied to the system reduce the ratio f/Ψ , the relative strength of vertical mixing, f , to the MOC, Ψ . As we will see in chapter 5, the vertical mixing f also includes the strength of the bottom overturning cell. In the case of North Atlantic wind stress, the bottom cell weakens as a result of increased stratification associated with stronger heat uptake and more productive Northern hemisphere biology (see Fig. 4.17e). While the reduction in bottom overturning cell may not be obvious in Fig. 4.15b, the time series of the bottom cell strength for any experiment shows a gradual reduction after an initial adjustment period. Since increased North Atlantic winds also increase the MOC strength, the ratio f/Ψ decreases with wind stress, pointing towards a negative dynamical atmospheric CO_2 feedback.

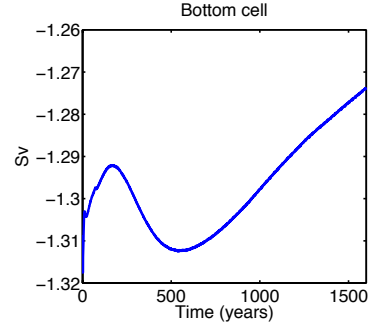


Figure 4.21: Time series of the minimum in the overturning stream function below 2500m depth in the Southern hemisphere in the $2 \times \overline{\tau_{NA}}$ experiment.

In the MITgcm experiments, the weakening of the bottom cell causes a build up of biological carbon which lowers atmospheric CO_2 levels several centuries after the wind stress step perturbation. As a result, regenerated carbon is trapped in the weakened bottom cell leading a negative atmospheric CO_2 feedback, consistent with the feedback framework outlined in section 2.4.2. We can compare the magnitude of the negative feedback associated with weakening bottom cell to the negative feedback associated with

carbon re-distribution and find that they are both of equal magnitude for all wind stress perturbations (estimated by calculating the terms in eqn. 2.36). Together, they act to reduce the atmospheric $p\text{CO}_2$ feedback to North Atlantic wind stress.

Overall, we find that the North Atlantic ocean $p\text{CO}_2$ is set by zonal advection, the radiative flux (buoyancy mixing) and vertical mixing (mechanical mixing), as opposed to mainly vertical mixing in the Southern Ocean highlighting the different dominant dynamical processes in both regions. We also find that the sensitivity of the atmospheric $p\text{CO}_2$ to wind stress is much less in the North Atlantic than in the Southern Ocean. The projected positive trend in the NAM is less than half as strong as for the SAM, so in terms of the future climate, we expect the NAM to only play a small role in the global carbon-climate feedback.

4.3 Discussion and summary

The goal of this section is to quantify the non-local equilibrium climate feedback of Southern Ocean wind stress on atmospheric $p\text{CO}_2$, $p\text{CO}_2^a$, using idealised coarse-resolution MIT-gcm experiments. It is found that wind stress-induced upwelling strengthens the export of nutrients North of the Southern Ocean by increasing the nutrient stream. As a result, area-mean equatorial biological productivity $\overline{F_{bio,eq}}$ increases. Due to stoichiometric ratio in the carbonate chemistry, the increased biological export from the surface to depth increases the Revelle Buffer Factor, R , in the Eastern tropical shadow zone/oxygen minimum zone located between 100 – 700m depth. Via the interior transport underneath the thermocline, the waters enriched in R are entrained into the nutrient stream and transported to the Northern sub-polar gyre where they are entrained into the seasonal thermocline, increasing the sub-polar gyre buffer factor and ocean $p\text{CO}_2$, $p\text{CO}_2^o$.

Counteracting the rise in R is an increase in local North Atlantic biology $\overline{F_{bio,na}}$ as a result of the stronger nutrient stream, which reduces the North Atlantic buffer factor. The overall change in mean buffer factor, $\Delta\overline{R_{na}}$ is therefore proportional to the change in the ratio $\frac{\overline{F_{bio,eq}}}{\overline{F_{bio,na}}}$. Due to the fairly linear relationship between atmospheric and oceanic $p\text{CO}_2$, $\Delta p\text{CO}_2^a$ also scales linearly with $\Delta\overline{R_{na}}$. We show that the effect is due to the adjustment of the ocean biogeochemistry by perturbing the nutrient- and light- half saturation

constant to alter the ratio of North Atlantic to Equatorial biology which leads to similar relationships between ΔpCO_2^o and $\Delta \overline{R_{na}}$.

The resulting atmospheric feedback is of roughly the same order as the local Southern Ocean effect, measured by the non-local feedback parameter ϵ_f , which is modelled to have a value of 0.8 – 1.1. The non-local feedback of Southern Ocean wind stress perturbations therefore magnifies the local feedback on atmospheric pCO_2 by 80% to 110%. The associated change in the North Atlantic buffer factor therefore constitutes a significant climate feedback that cannot be measured by considering changes in the Southern Ocean carbon sink alone. The impact of Southern Ocean wind stress is therefore more significant than concluded at the end of section 3.1. Similar mechanisms are seen in fully coupled global models such HadGEM2-ES, though further study with high-resolution coupled atmosphere-ocean GCMs is needed to fully assess the impact in the modern climate.

Compared to the local atmospheric carbon feedback of the North Atlantic sub-polar gyre wind, which is set by zonal advection, the radiative flux and vertical mixing contributions to local $\overline{pCO_2^o}$, we find that Southern Ocean winds modulate the North Atlantic carbon flux more than the local winds. For example, a Southern Ocean wind stress perturbation of $\Delta \overline{\tau_{SO}} = 0.13 \text{ N/m}^2$ results in an atmospheric pCO_2 feedback of roughly 20ppm (without the atmospheric pCO_2 change due to the local Southern Ocean feedback, see Fig. 4.9a). The change in atmospheric pCO_2 due to a perturbation in North Atlantic winds of the same size however, causes a 10ppm increase in atmospheric pCO_2 (Fig. 4.20b). The same relative difference is seen for all magnitudes of wind stress perturbations. The effect of Southern Ocean winds on the North Atlantic carbon sink is therefore twice as strong as the influence of the local winds.

Caveats of the proposed mechanism include the failure of models to accurately represent the extent of the oxygen-minimum zones. A phenomenon coined “nutrient trapping” [Najjar, 1990, Aumont et al., 1999] results in overestimation of the strength of the oxygen minimum in the Tropics. This is because the upwelled nutrients are used in the mixed layer and rained down in the same region, causing an exaggerated self-sustaining cycle, which is a result of the unresolved complexity in equatorial undercurrents. Nonetheless, the proposed mechanism should still be valid since oxygen minimum zones are observed in the ocean [Stramma et al., 2009].

A second caveat is the use of a coarse-resolution model. However, we do not expect the values of ϵ_f to change with resolution since both Southern Ocean pCO_2^a and the relative response of the North Atlantic and Equatorial biology are similarly dependent on the wind-induced upwelling in the Southern Ocean. Thirdly, Redfield ratios are constant both spatially and temporally in the model, which is not the case in the real world [Weber and Deutsch, 2010], though how variations in the Redfield ratio will affect the result is unknown. With regards to our particular model set-up, we noted in section 2.4.1 that the biological productivity in our model is relatively weak. However, the mechanism that we describe depends on the relative strength of the equatorial and the North Atlantic sub-polar gyre biology, not on the total biological activity. For systems with higher global mean biological activity, we expect the atmospheric CO_2 response to be larger since there is a larger biological reservoir of carbon to act as a source for the atmospheric CO_2 changes (e.g. Ito and Follows [2005], see section 5.2.4). Therefore, we believe the magnitude of the non-local feedback to be a lower estimate.

Lastly, the size of the sector model plays an important role in setting the timescale of the non-local Southern Ocean winds feedback on pCO_2^a . The feedback timescale depends on basin-wide advection: the Buffer Factor anomaly in the shadow zone has to be advected zonally from East to West before it is advected northwards by the nutrient stream. In a wider basin therefore, the cross-basin zonal advection timescale is longer. To illustrate this dependence of the timescale on basin width, an identical MITgcm set up 40° in longitude wide is used. Figure 4.22 shows the time evolution of atmospheric, Southern Ocean and

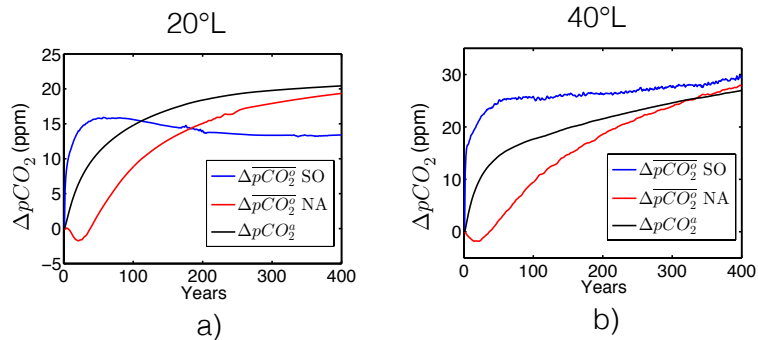


Figure 4.22: Time evolution of ΔpCO_2^a (black), North Atlantic $\Delta \overline{pCO_2^a}$ (red) and Southern Ocean $\Delta \overline{pCO_2^a}$ (blue) in a $2 \times \tau_{SO}$ for a) the 20° longitude wide model and b) the 40° longitude wide model.

North Atlantic sub-polar gyre pCO_2 anomalies in $2 \times \tau_{SO}$ experiment for both the 20° and

40° wide models. While the non-local signal appears in the North Atlantic $p\text{CO}_2$ after 20 years in each case, the equilibration timescales varies. After 500 years, atmospheric $p\text{CO}_2$ has equilibrated in the 20° model while equilibration takes twice as long in the 40° longitude model. However, the strength of the non-local feedback ϵ_f is similar. The width of the basin therefore determines the timescale of the non-local feedback but not the magnitude. Similarly, the zonal currents in the model are relatively weak compared to observations (see section 2.3.3), however we believe the strength of the currents affects only the advective timescales in the model. The weak currents therefore affect the timescale of the non-local feedback but not the magnitude.

The lack of Pacific and Indian Oceans could also affect the net non-local feedback. However, the Pacific and Indian Oceans do not constitute a large sink of atmospheric carbon so are unlikely to result in similar feedbacks as seen in the Atlantic. As atmospheric $p\text{CO}_2$ increases, Pacific and Indian Ocean $p\text{CO}_2$ is expected to track atmospheric $p\text{CO}_2$. Without detailed study of the dynamical and biogeochemical changes induced by increased Southern Ocean winds in the global ocean however, the exact contribution of the Pacific and Indian Oceans is difficult to estimate. Further study is therefore required to determine the exact equilibration timescale and magnitude in a realistic bathymetry domain.

The relationship between the North Atlantic buffer factor and equatorial biology has further implications for the climate as well, since the ratio of North Atlantic to Equatorial biology can be altered by many processes such as stratification, cloud cover, other sources of macro-nutrients, iron supply, to name a few. Our results show the influence of oxygen-minimum zones on North Atlantic surface chemistry. Given recent results from studies that show expansion of tropical oxygen-minimum zones [Stramma et al., 2012], a thorough understanding of the biogeochemical drivers of the North Atlantic Buffer Factor is therefore needed. The North Atlantic sub-polar region constitutes one of the main sinks for anthropogenic CO_2 and our results show that changes in the buffer factor of as small as 0.5 could result in a positive atmospheric carbon feedback of 10 – 30ppm.

Climate feedback of surface warming on atmospheric $p\text{CO}_2$

Increased global mean temperatures are one of the most studied and discussed effects of anthropogenic CO_2 -induced climate change [Manabe and Stouffer, 1994, Rogelj et al., 2012, Gillett et al., 2013]. The subsequent effect of increased ocean surface temperatures on the ocean carbon cycle constitutes one of the pre-eminent climate-carbon feedbacks within the climate system [Bacastow, 1996, Joos et al., 1999, Gregory et al., 2003, Frolicher et al., 2014], due to both solubility and dynamically-induced carbon feedbacks [Goodwin et al., 2009]. As temperature increases, the solubility of gasses is reduced which affects surface properties such as ocean $p\text{CO}_2$. Since ocean $p\text{CO}_2$ sets a large part of the air-sea flux, an increase in temperatures could have a strong regional influence on the air-sea flux of carbon, in particular where the carbon uptake is large such as in the high-latitudes [Sarmiento et al., 1998, Perez et al., 2013]. The first aim of this chapter is therefore to deduce the role of solubility on high latitude carbon uptake in climate change scenarios.

Globally, an increase in ocean heat uptake affects the solubility of carbon throughout the ocean basin. However, warming the surface ocean also affects the stratification and large scale circulation. Stratification and circulation are key factors in the ocean carbon cycle response to forcing, alongside the temperature-induced reduction in gas solubility [Sarmiento et al., 1998, Takahashi et al., 2003, Bernardello et al., 2014]. Studies such as Goodwin et al. [2009] quantify the role of the increase in ocean heat content over the period 1961-2003 on ocean carbon using box models and an Earth System Model. They suggest that buoyancy-induced changes in the overturning circulation could have an equally important effect on atmospheric CO_2 as solubility. Studies such as Joos et al. [1999] examine competing changes in saturated carbon and regenerated carbon due to warming-induced changes in the large scale dynamics, on a 500-year timescale. As ocean surface temperatures increase, changes in dynamics will also affect disequilibrium carbon (defined in section 1.2.2), which depends directly on the large scale circulation. Our second goal is therefore to quantify the global ocean carbon response to surface warming, with

a particular focus on the response of saturated, regenerated and disequilibrium carbon to warming-induced changes in solubility versus dynamics.

This chapter will therefore quantify the climate feedback of surface warming on atmospheric $p\text{CO}_2$. In section 5.1, the MITgcm carbon cycle to surface warming perturbations is described, after which the role of surface warming on high latitude carbon is determined. The key findings discussed in detail in section 5.1 are summarized below:

- In the Southern Ocean, $p\text{CO}_2^o$ anomalies due to warming can be predicted by calculating the change in solubility of CO_2 , since local DIC remains constant;
- In the North Atlantic sub-polar gyre, the $p\text{CO}_2^o$ is found to vary little with warming;
- Polar amplification of Southern or Northern temperatures is found to cause opposing changes in the North Atlantic and Southern Ocean $p\text{CO}_2^o$.

Afterwards, section 5.2 will discuss surface warming-induced changes of the large scale circulation and how they affect the different carbon reservoirs by using MITgcm surface warming and salinity perturbation experiments, as well as box models. The key findings discussed in details in section 5.2 are summarized below:

- Over 400 a year period, warming-induced dynamical changes have little effect on the atmospheric $p\text{CO}_2$ due to compensation of saturated and disequilibrium carbon reservoirs;
- From 400 to 3000 years, polar amplification of Northern and Southern temperatures results in a reduction in atmospheric $p\text{CO}_2$ due to isolation of regenerated carbon at depth. The increase in regenerated carbon compensates the loss of saturated carbon.
- Salinity perturbation experiments are used to shown that changes in the MOC have a much smaller effect of atmospheric $p\text{CO}_2$ than changes in the bottom cell due to different degrees of carbon compensation.
- Simple box models for preformed and regenerated carbon are constructed to predict atmospheric $p\text{CO}_2$ anomalies due to changes in the MOC and bottom cell.

Lastly, we conclude the chapter in section 5.3 through:

- Discussion of the implications of the warming-induced ocean carbon response for anthropogenic ocean carbon uptake;

- Discussion of the implication of natural carbon compensation in warming scenarios;
- Discussion of the caveats in our study.

5.1 High latitude warming

5.1.1 Description of the model response to surface warming

In this section, the response of the MITgcm sector model to perturbations in the surface radiative forcing is discussed for the multiplicative base warming scenario (see section 2.3.4), which ensures that the North Atlantic sub-polar region is warmed more than the Southern channel [Xie et al., 2010], while drawing specific examples from the $1.2 \times T$ experiment (see Fig. 5.1). We analyze the MITgcm response to warming after 400 years since the large scale circulation equilibrates over this time.

As a result of the warming imposed, the wind-induced gyre circulation advects the tropical temperature anomaly along western boundary currents. The density anomalies in the model are predominantly temperature driven so that surface density anomalies show a similar pattern to the temperature anomalies. To maintain a pole-to-pole overturning circulation, mode waters formed in the Northern high latitudes must be able to travel along isopycnals that outcrop in the Southern channel, so there must be a range of densities shared between the two poles [Wolfe and Cessi, 2010]. Since the North Atlantic warms more than the Southern channel, the shared density range between the South and North high latitude is altered, which drives changes in the

overturning circulation. Within 10 years of the perturbation, the MOC decline is on the order of 1 Sverdrup in the $1.2 \times T$ base warming experiment, corresponding to 25% of the control mean. After 400 years however, the MOC recovers and overshoots such that its long-term strength after the perturbation is higher than in the unperturbed control state.

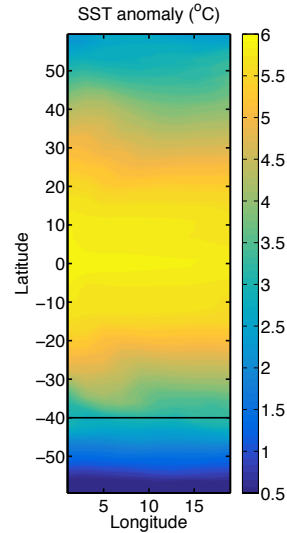


Figure 5.1: Sea surface anomaly in MITgcm for the $1.2 \times T$ base warming perturbation scenario.

We define the global mean sea surface temperature T_G as the mean of the ocean temperature between 0-20°E, 60°S to 60°N at the ocean’s surface, and the global mean SST anomaly ΔT_G as the difference between T_G at a given time in one of the perturbed experiments and T_G in the control integration. Figure 5.2a then shows the MOC strength as a function of ΔT_G after 400 years, illustrating the increasing trend. *Note that in this section, anomalies are defined as averages over the last 25 years of a 400 year-long run under a perturbed scenario minus the control run mean state unless stated otherwise.* We find that the response of the MOC in the model is due to a local increase in convection in the North Atlantic and does not translate into an increase in the pole-to-pole overturning circulation.

The overall increase in the MOC strength with ΔT_G results from a stronger meridional temperature gradient since the equator is warmed more than the Northern high latitudes (see Fig. 5.1). The stronger meridional temperature gradient causes stronger cooling of northward advected water parcels that are exchanged between the sub-tropical and sub-polar gyre. The increase in cooling rate results in stronger buoyancy forcing and therefore increased local convection in the sub-polar gyre [e.g. Stommel, 1961, Schloesser et al., 2012]. However, the NADW mode water has a higher buoyancy than in the control state so that the depth of the MOC is reduced (Fig. 5.2b). The stronger warming in the North Atlantic relative to the Southern channel leads to a mismatch in density and an overall reduction of the pole-to-pole overturning. As a result, the North Atlantic becomes more decoupled from the Southern Ocean: more of the North Atlantic mode water is upwelled in the interior before it reaches the Southern Ocean. The schematic in Fig. 5.3a shows the density structure required for a pole-to-pole overturning. As the North Atlantic convection increases due to stronger air-sea surface heat fluxes, there is an increase in the North Atlantic diffusive MOC (also shown in Fig. 5.3), whereby buoyancy forcing in the interior occurs through vertical mixing.

Figure 5.3b shows the zonally-averaged residual meridional overturning streamfunction Ψ anomaly for the $1.2 \times T$ perturbation experiment. The figure shows the increase in the diffusive MOC and reduction of the pole-to-pole overturning. The decrease in the pole-to-pole MOC can be measured by calculating the northward Ekman transport at 40°S, defined here as the volume transport integrated between 0 – 20°E and 0-100m depth and shown in Fig. 5.2c. There is also a weakening of the bottom water cell due an increase in

surface stratification in the Southern channel inhibiting the formation of Antarctic Bottom Water.

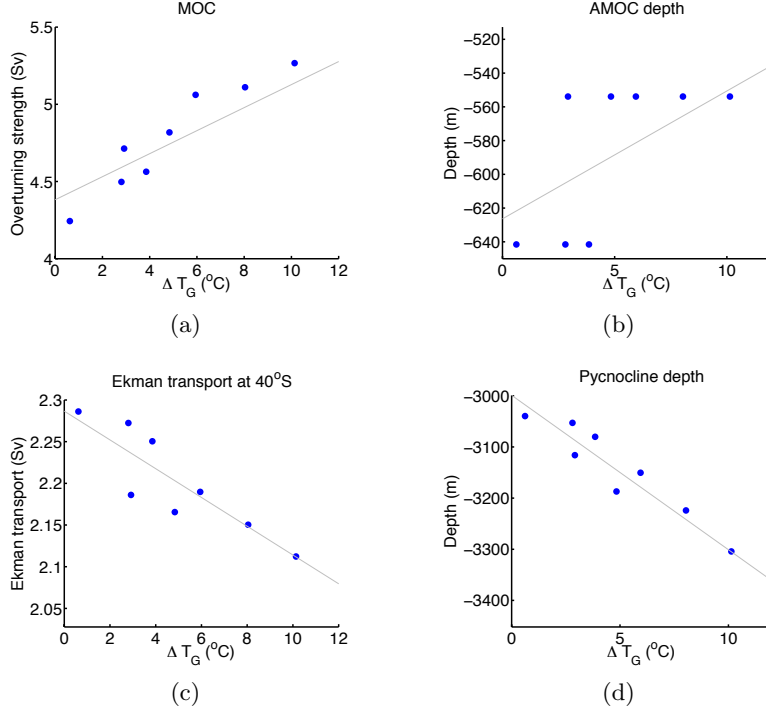


Figure 5.2: MITgcm diagnostics as a function of global mean SST anomaly, ΔT_G , after 400 years of the base warming scenario: a) Maximum of the meridional overturning stream function between 350-1550m depth and 30°-60°N; b) Depth of MOC maximum; c) Ekman transport at 40°S between 0 – 100m depth; d) Pycnocline depth calculated using eqn. 3.1 as in Gnanadesikan [1999]. Grey lines show a least-squares linear fit.

The surface temperature anomaly propagates into the basin interior (away from the surface) through the subduction of mode waters formed North of the channel and convection in the Northern sub-polar gyre. Figure 5.4a shows the zonally averaged temperature anomaly resulting from a $1.2 \times T$ perturbation. The positive temperature anomalies are strongest at the surface but are advected into the water column just North of the channel and in the Northern sub-polar gyre. Due to a weakening of the bottom water cell, penetration of the warming anomaly to depths below 2500m is limited. The mixed layer temperature anomaly (Fig. 5.4b) is an imprint of the perturbation forcing profile (discussed in section 2.3.4) and the advection of the tropical anomaly along western boundary currents of the sub-tropical gyres. The increase in buoyancy in the upper ocean (Figs. 5.4a-b) associated with the warming depresses isopycnals and causes a deepening of the pycnocline. Therefore, as warming increases, the pycnocline deepens (Fig. 5.2d).

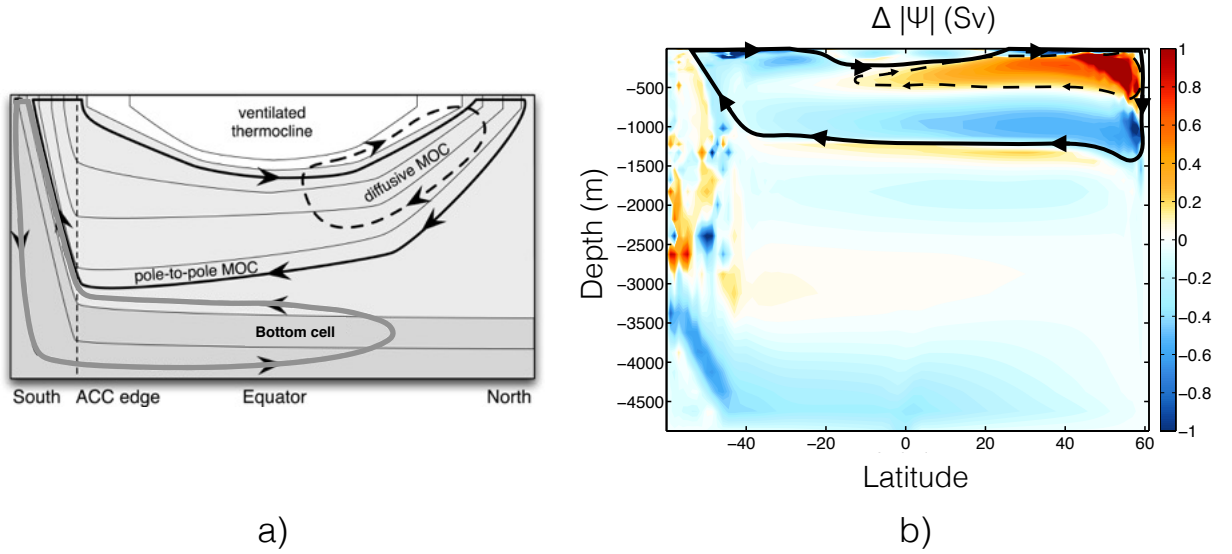


Figure 5.3: a) Diagram of the MOC (solid black line) with depth and latitude in a rectangular idealized basin from Wolfe and Cessi [2010], thin grey lines are isopycnals, and the diffusive overturning is shown in dashed lines. The bottom cell is shown in a thick grey line. b) Zonally-averaged anomaly in the magnitude of the residual stream function in the $1.2 \times T$ MITgcm base warming experiment, where the pole-to-pole overturning is marked in solid lines and the diffusive overturning in dashed lines.

The pycnocline depth determines the carbon concentration of upwelling waters since the carbon content increases with depth. A deepening pycnocline therefore results in a flushing of carbon from the deep ocean to the surface and atmosphere [Lauderdale et al., 2013, Munday et al., 2014]. Figure 5.4c shows the dissolved inorganic carbon anomaly for the $1.2 \times T$ perturbation scenario. In the upper overturning cell, a reduction in the solubility of carbon due to warming results in a decrease in carbon concentrations. Due to shallowing of the upper cell and subsequent isolation¹ of underlying waters, carbon builds up at 1500m-depth in the Northern hemisphere. An increase in carbon is not seen in the Southern hemisphere due to deepening of the isopycnals such that water from 1500m-depth is ventilated in the Southern channel. The ventilation of deeper, more carbon-rich waters in the Southern channel results in a smaller negative mixed layer *DIC* anomaly than elsewhere in the basin, as seen in Fig. 5.4d, offsetting to some extent the effect of a reduced residual circulation. In the bottom water cell, a reduction in circulation results in a build up of dissolved inorganic carbon. Figures 5.4e and 5.4f show the zonally averaged preformed and regenerated anomalies of *DIC*, respectively. The signal throughout the basin is found to be dictated by the preformed component in the upper 1000m. In the upper cell, the reduction in solubility as a result of warming causes a decrease in pre-

¹Where ‘isolation’ refers to the reduced ventilation of deep waters as the MOC shallows and weakens.

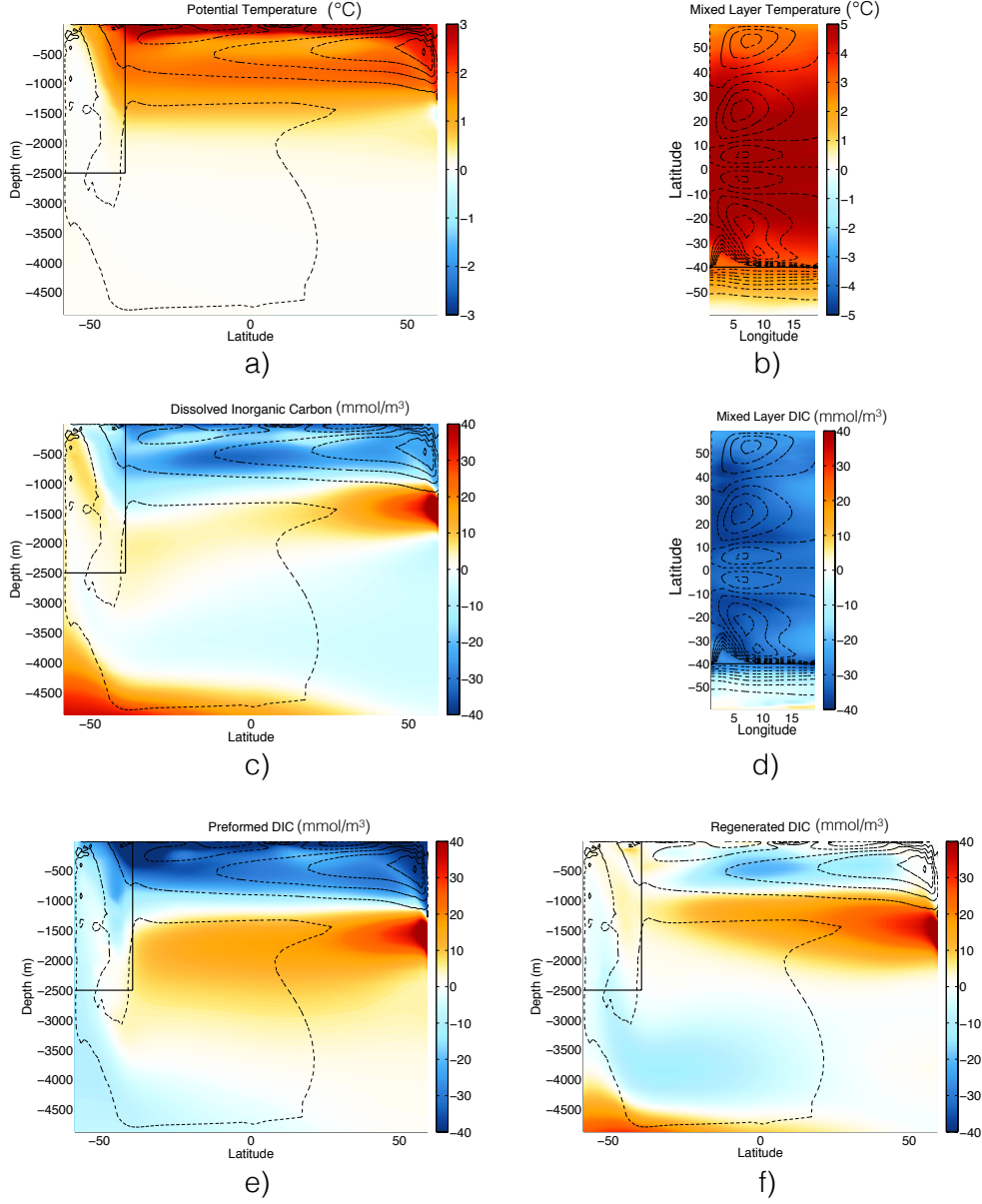


Figure 5.4: Anomalies in the $1.2 \times T$ base warming experiment: a) zonal mean and b) mixed layer potential temperature, c) zonal mean and d) mixed layer DIC, e) zonal mean preformed carbon C_{pre} and f) zonal mean regenerated carbon C_{reg} . Dashed lines in the zonal mean plots (a,c,e,f) denote the perturbed residual overturning stream function contours spaced 1 Sv-apart. Dashed lines in the mixed layer plots (b,d) show contours of the perturbed barotropic stream function.

formed carbon. In the bottom water cell, the shallowing of the Northern branch of the overturning circulation and decrease in bottom water formation in the Southern channel isolates preformed carbon in the expanding bottom water cell. Global surface biological productivity is reduced due to both increased stratification and a reduction in Southern ocean residual circulation which reduces the upwelling of nutrients.

The reduction in biological activity is notably larger in the Southern hemisphere mid-latitudes, North of the channel, where the biology is predominantly nutrient limited such that the reduction in residual circulation has a particularly strong effect. The weaker biological productivity causes a reduction of regenerated carbon in the upper overturning cell. Below 1000m, the shallowing of the upper overturning cell and weakening of the bottom water cell results in accumulation of regenerated carbon over time. Importantly, regardless of the shallowing of the Northern section of the overturning, the deepening of isopycnals in the Southern hemisphere causes the positive regenerated carbon anomaly to be upwelled in the Southern channel. The deepening of isopycnals North of the channel therefore acts as a way to expose the accumulated regenerated carbon to the surface. In all temperature perturbation experiments, the patterns of preformed and regenerated carbon anomalies are dictated by the response of the upper and lower overturning cells, the solubility and the deepening of the isopycnals.

Figure 5.5 shows the change in atmospheric $p\text{CO}_2$, $\Delta p\text{CO}_2^a$, as a function of the global anomaly in sea surface temperature ΔT_G as modelled in the MITgcm sector model. For all anomalies ΔT_G , the net result is an increase in atmospheric CO_2 as shown in a host of other studies [Sarmiento et al., 1998, Goodwin and Lenton, 2009]. Studies such as Joos et al. [1999] attribute the majority of the atmospheric CO_2 signal to solubility. Our aim is therefore to deduce the role of solubility in the high latitudes in setting the atmospheric CO_2 perturbations, since the high latitudes form key areas for ocean carbon uptake.

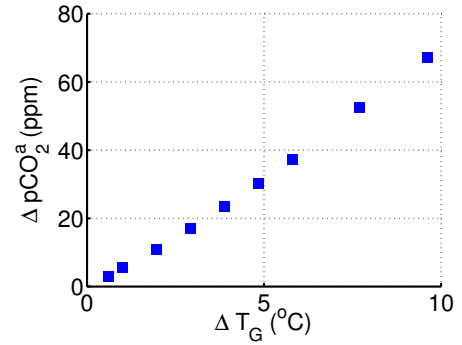


Figure 5.5: Change in atmospheric CO_2 in MITgcm base warming experiments as a function of global sea surface temperature anomaly ΔT_G .

5.1.2 Southern Ocean surface carbon

The response of the local ocean $p\text{CO}_2$, $p\text{CO}_2^o$, to a change in temperature T can be quantified using the empirical relationship determined by Takahashi et al. [1993]:

$$\frac{\partial p\text{CO}_2^o}{\partial T} = 0.0423 p\text{CO}_2^o. \quad (5.1)$$

The relationship quantifies the reduction in solubility of dissolved CO_2 gas in water as the temperature increases, as long as DIC concentrations remain constant. The uncertainty in the variation of $p\text{CO}_2^o$ with T is quoted as $\frac{\partial p\text{CO}_2^o}{\partial T} = (0.0423 \pm 0.0002^\circ) p\text{CO}_2^o$ over a range of $2.1^\circ - 24.5^\circ\text{C}$, so it holds well for the high latitude ocean². Equation 5.1 is widely used when separating solubility vs. other (dynamical or biological) contributions in ocean $p\text{CO}_2$ variations [Takahashi et al., 1993, 2009, McKinley et al., 2011, Roedenbeck et al., 2013, Wanninkhof et al., 2013, Majkut et al., 2014]. To estimate the change in Southern Ocean $p\text{CO}_2^o$ relative to the control mean, $\Delta \overline{p\text{CO}_2^o}$, as a result of a sea surface temperature perturbation in the Southern Ocean $\Delta \overline{T}$, eqn. 5.1 can be re-written as:

$$\Delta \overline{p\text{CO}_2^o} = \langle \overline{p\text{CO}_2^o} \rangle 0.0423 \Delta \overline{T}, \quad (5.2)$$

where $\langle \dots \rangle$ indicates the control mean and the overline indicates the area average in the Southern Ocean (namely, across all longitudes and between $60^\circ - 40^\circ\text{S}$ for the Southern channel). Equation 5.2 allows us to disentangle the effects of solubility from other dynamical influences (although there will be a feedback of the dynamics on temperature). Figure 5.6 shows that the modelled response of the Southern channel $\Delta \overline{p\text{CO}_2^o}$ (squares) follows the predicted trend (solid line) based on solubility, given by eqn. 5.2. The error between the prediction and the modelled response is less than 5% in all cases, such that the response appears to be determined predominantly by the solubility trend, although this is only valid if DIC concentrations remain constant.

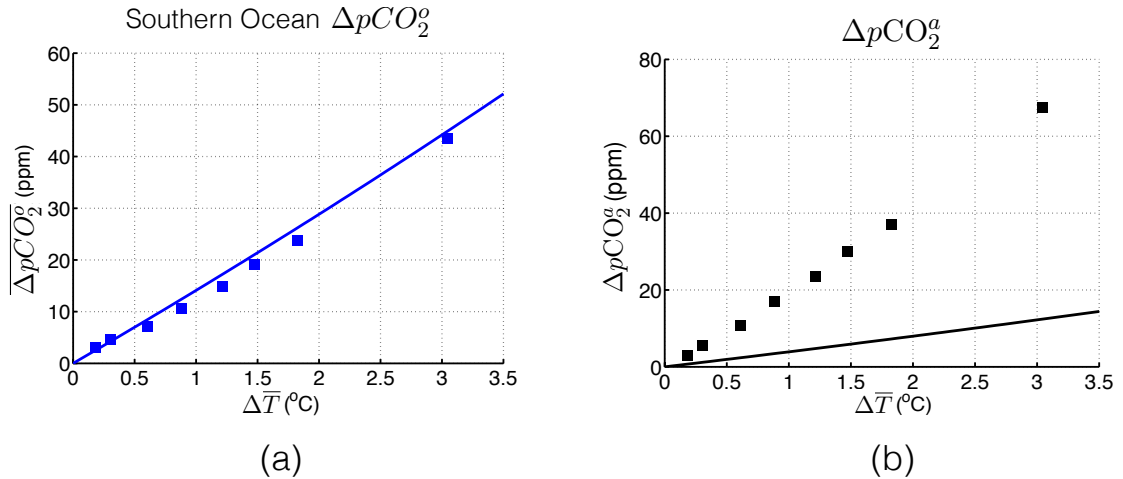


Figure 5.6: a) Change in $\overline{p\text{CO}_2^o}$, in ppm, as a function of Southern channel SST anomaly $\Delta \overline{T}$. The squares show MITgcm data from base warming experiments and the solid blue line is the prediction based solely on solubility effects (eqn. 5.2). b) Predicted $\Delta p\text{CO}_2^a$ in response to Southern Ocean warming using eqn. 5.3 (solid black line) and $\Delta p\text{CO}_2^a$ MITgcm output from temperature perturbation experiments (black squares).

²Equation 5.1 wouldn't hold in the tropics, where sea surface temperatures are above 24.5°C

When a positive temperature perturbation is applied to the ocean, it will result in an overall increase in buoyancy and deepening of the pycnocline. As the pycnocline deepens, the depth from which water is upwelled therefore deepens and exposes more carbon rich water to the surface [Jones et al., 2011, Lauderdale et al., 2013]. As the residual circulation in the Southern Ocean weakens, the volume of water supplied reduces. However, the deepening of the pycnocline ensures that the supply of carbon is increased in order to balance the increased surface outgassing, resulting in a small change in DIC (e.g. Fig. 5.4d show reductions in DIC between 0 to 10 mmol/m³ in the Southern channel compared to reductions in DIC of 30 to 40 mmol/m³ elsewhere in the basin). To match the deepening of isopycnals and Southern Ocean carbon, we can define $\Delta_x b_h$ as

the horizontal density window that is outcropped in the Southern channel. The window $\Delta_x b_h$ can be mapped on a vertical density window $\Delta_z b_v$ in the ocean interior since tracer transport away from the surface occurs predominantly along isopycnals [Gnanadesikan, 1999, Wolfe and Cessi, 2010] (see Fig. 2.11a.).³ Figure 5.7 shows the evolution of the depth of the density window $\Delta_z b_v$ as well as the evolution of $\Delta \overline{pCO_2^o}$.

After the initial few years, the Southern channel pCO_2^o anomaly grows with a similar timescale to the deepening of the density window Δb_v . The maximum $\Delta \overline{pCO_2^o}$ is achieved between 100 – 150 years after the initial perturbation. The maximum $\Delta \overline{pCO_2^o}$ follows eqn. 5.2 to within a few percent. The maximum $\Delta \overline{pCO_2^o}$ is therefore set by the solubility but the timescale is set by the adjustment of the density structure. Following the maximum in $\Delta \overline{pCO_2^o}$, there is a slow decline associated with accumulation of regenerated carbon in the weakening bottom circulation cell [Schmittner, 2005, Schmittner et al., 2007], which

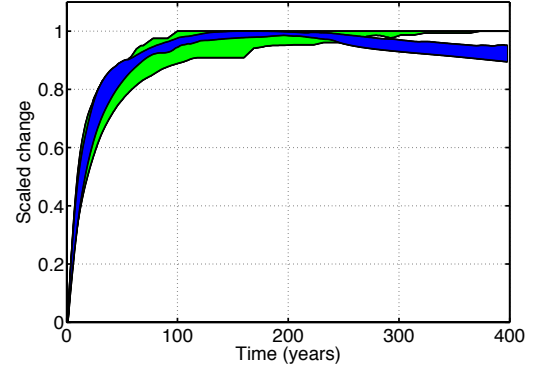


Figure 5.7: Envelopes of the time evolution of the depth of the density window $\Delta_z b_v$ (green) and Southern Ocean ΔpCO_2^o (blue), scaled such that the maximum value of each curve is 1, for all temperature restoring perturbation experiments. The envelopes show the spread of the evolution of each quantity over all the perturbation experiments.

³For example, Fig. 2.10 shows that the range of isopycnals outcropped in the Southern Ocean $\Delta_x b_h$ corresponds to a vertical density window in the ocean interior $\Delta_z b_v$ between 1500m and 300m depth.

is not captured by the deepening timescale of the density window since the equilibration timescale of regenerated carbon is much longer.

In general, $\Delta \overline{pCO_2^o}$ follows the solubility trend well. We can use the $\Delta \overline{pCO_2^o}$ to estimate the associated change in ΔpCO_2^a using eqn. 2.42 (derived in section 2.4.2):

$$\Delta pCO_2^a = P_{SO} \Delta \overline{pCO_2^o}, \quad (5.3)$$

where P_{SO} is the area fraction of the Southern Ocean weighted by wind stress. Figure 5.6b shows the ΔpCO_2^a estimated using eqn. 5.3 and the MITgcm data. The figure shows that the response modelled by MITgcm is much larger than the response predicted by the scaling, meaning that the natural carbon ΔpCO_2^a is set by processes other than the Southern Ocean carbon anomaly, the details of which will be examined in section 5.2.

While the results use an idealised ocean model, we believe the mechanisms discussed are relevant for the real ocean. The key dynamical difference is the presence of eddies in the real ocean, but since $\Delta \overline{pCO_2^o}$ depends largely on solubility and deepening of the pycnocline due to increased upper ocean buoyancy, we do not expect that the presence of eddies will drastically change the results. However, they will affect the timescales over which $\Delta \overline{pCO_2^o}$ adjusts [Gent and McWilliams, 1990, Allison et al., 2011].

5.1.3 North Atlantic surface carbon

In the North Atlantic sub-polar gyre, similarly to the Southern channel, we can use eqn. 5.2 to estimate the local change in ocean pCO_2 in the sub-polar gyre, $\Delta \overline{pCO_2^o}$, in response to warming (for the North Atlantic, the overline indicates the spatial average between $0^\circ - 20^\circ E$ and $40^\circ - 60^\circ N$). Figure 5.8 shows the predicted $\Delta \overline{pCO_2^o}$ (solid thick line) and the MITgcm output (squares).

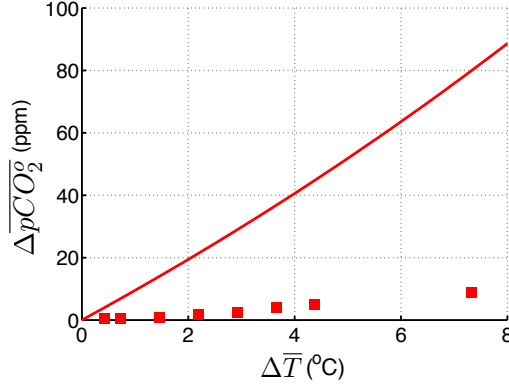


Figure 5.8: Change in North Atlantic pCO_2^o , as a function of North Atlantic sub-polar gyre sea surface temperature anomaly $\Delta \bar{T}$. The squares show MITgcm data from the base warming experiments and the solid red line is the prediction based solely on solubility effects (eqn. 5.2).

Unlike the Southern channel, the North Atlantic ΔpCO_2^o does not follow the solubility-predicted trend. The $\Delta \overline{pCO_2^o}$ is negligible for local temperature changes up to $\Delta \bar{T} = 2^\circ\text{C}$: the increase in $\Delta \overline{pCO_2^o}$ due to the increased temperature via solubility is balanced by dynamical effects induced by surface buoyancy forcing. The negative dynamical effects result from the loss of DIC in the upper ocean (0-1000m) due to solubility (see Fig. 5.4c), since North Atlantic surface waters are sourced from intermediate mode waters more affected by the surface warming, as opposed to deeper waters in the Southern Ocean. Past a threshold of $\Delta \bar{T} = 2^\circ\text{C}$, the change in $\overline{pCO_2^o}$ is only 10% of the predicted value due to solubility changes. The result suggests that the North Atlantic surface pCO_2^o remains largely constant with surface warming perturbations. Using the North Atlantic ΔpCO_2^o to predict the atmospheric ΔpCO_2^a response results in a large mismatch since the predicted ΔpCO_2^a is negligible (not shown).

5.1.4 High latitude carbon response to polar warming

To test the robustness of our results we can perturb the dynamics of the system by repeating the warming experiments with a different set of forcing profiles. In section 2.3.3, the base warming profile is one where the percentage warming at the Northern pole is twice that at the equator.

There is evidence of strong warming in the Arctic in the HadGEM2-ES $2\times CO_2$ and $4\times CO_2$ step experiments however, a phenomenon known as polar amplification [Haywood et al., 1997, Chapman and Walsh, 1993, Holland and Bitz, 2003, Bekryaev et al., 2010,

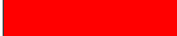
Experiment	Northern boundary temperature T_N	Southern boundary temperature T_S	colour code
Base	$2\beta - 1$	0	
$+T_N$	$4\beta - 3$	0	
$++T_N$	$6\beta - 5$	0	
$+T_S$	$2\beta - 1$	$2(\beta - 1)$	

Table 5.1: Table listing the restoring temperature perturbations to the Northern and Southern boundary temperatures in additional warming experiments, where β is the magnitude of the global temperature perturbation, so the $1.2xT$ experiment has $\beta = 1.2$, along with the respective colour codes used throughout this chapter.

Pithan and Mauritsen, 2014]. The polar warming can be seen in the Atlantic zonal mean temperature anomalies (Fig. 5.9a), where the Northern pole warms more than other latitudes and the polar amplification intensifies with the strength of the atmospheric CO_2 perturbation. We can use those profiles to construct a second and third set of experiments whereby the Northern boundary temperature T_N is multiplied by factors $4\beta - 3$ and $6\beta - 5$, where β is the magnitude of the global temperature perturbation, so the $1.2xT$ experiment has $\beta = 1.2$. The relative warming at the Northern boundary is therefore increased, to mimic different degrees of polar amplification. Similarly, we run a fourth set whereby the temperature restoring is perturbed like in the base warming scenario, but T_S is increased using $T_S = 2(\beta - 1)$, such that $T_S = 1^\circ\text{C}$ in the $1.5 \times T$ experiment, modelling milder Southern Ocean warming, but enough to trigger a shutdown of the bottom overturning cell [Manabe et al., 1990, Xie et al., 2010, Majkut et al., 2014]. The experiments will be referred to as the “ $+T_N$ ”, “ $++T_N$ ” and “ $+T_S$ ” experiments respectively and assigned a specific colour which will be used throughout this chapter to plot results from the corresponding experiments. The experiments and colour codes are summarised in table 5.1. To illustrate the effect of the different warming scenarios, Fig. 5.9 shows an example set of the perturbation in the warming restoring profiles $\Delta\theta^c$ for the different sensitivity experiments for the $1.2 \times T$ experiment.

The significance of the boundary conditions is most relevant for convection at high latitudes and the strength of the resulting overturning circulation. Warming of the Northern temperature in the $+T_N$ and $++T_N$ experiments causes an increase in the stratification in the Northern high latitudes, without increasing the meridional temperature gradient to the same extent as in the base warming scenarios. The increase in stratification is dominant and causes a reduction in the MOC unlike in the base warming scenario. Similarly,

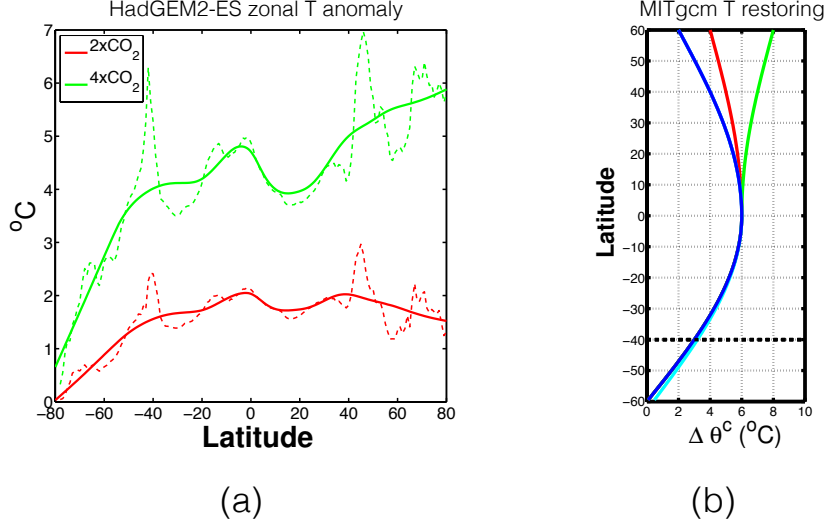


Figure 5.9: a) Atlantic zonal mean sea surface temperature anomalies in HadGEM2-ES at the end of the $2\times\text{CO}_2$ and $4\times\text{CO}_2$ step experiments relative to the pre-industrial value; b) Perturbations to the temperature restoring profile $\Delta\theta^c$ as a function of latitude for the base (blue), $+T_N$ (red), $++T_N$ (green) and $+T_S$ (cyan) warming scenarios in the $1.2\times T$ scenario.

a warmer Southern boundary T_S causes a reduction in the AABW bottom water cell. Figure 5.10 shows the response of both the MOC (panel a) and AABW (panel b) cells in the different sensitivity experiments. Note that the strength of the AABW bottom water cell is given as the absolute magnitude, since the streamfunction in the AABW bottom water cell is negative. While the MOC decreases with ΔT_G in the $+T_N$ experiment, the strong warming in the $++T_N$ experiment causes an effective shutdown of the circulation for a ΔT_G larger than 8°C . Similarly, the $+T_S$ experiment results in a stronger reduction in the AABW cell relative to the base scenario, such that the cell is shut down for a ΔT_G larger than 6°C .

The result of the dynamical changes induced by the perturbed temperature restoring boundary conditions on the Southern Ocean and North Atlantic $\overline{p\text{CO}_2}$ is shown in Fig. 5.11 along with the estimated changes due to solubility alone (using eqn. 5.2). In the Southern channel (Fig. 5.11b), an increased Northern or Southern boundary temperature results in a reduced $\overline{p\text{CO}_2}$ anomaly. However, we find that the positive effect of solubility on $\overline{p\text{CO}_2}$ is still generally stronger than the negative effect of dynamics. Only in the extreme Northern warming case, $++T_N$, are both of the effects roughly equal and opposite, but only for weak warming where $\Delta\overline{T}$ is less than 2° . Therefore, we find that the results from the previous section generally hold, except in cases of extreme of Northern warming.

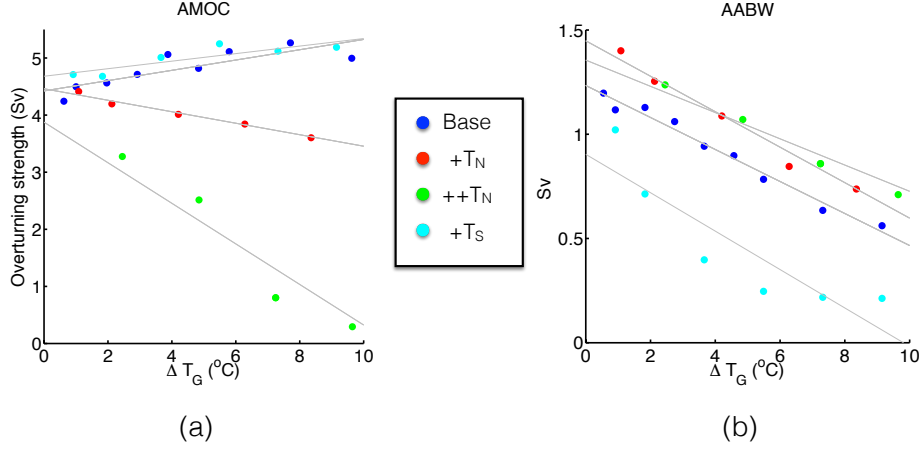


Figure 5.10: Strength of the a) MOC upper overturning cell and b) AABW lower overturning cell in the Northern and Southern boundary temperature sensitivity experiments as a function of the global temperature anomaly ΔT_G . The colours correspond to warming scenarios listed in table 5.1.

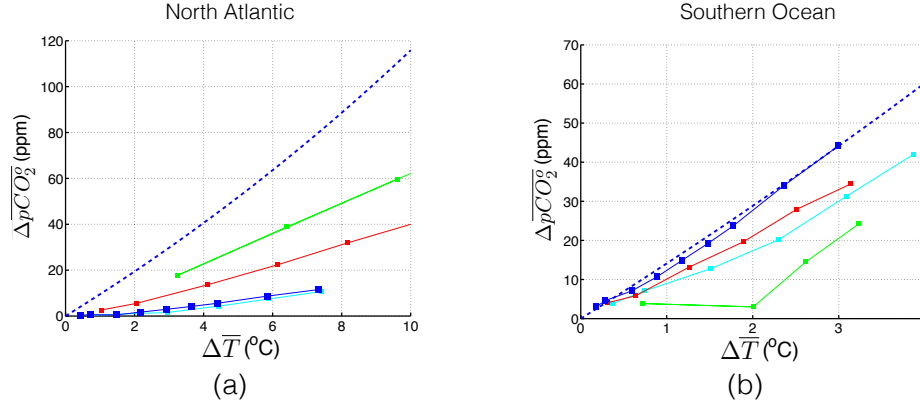


Figure 5.11: a) North Atlantic $\Delta \overline{pCO_2}$ and b) Southern Ocean $\Delta \overline{pCO_2}$ in response to a local $\Delta \overline{T}$. The colours correspond to the set of experiments listed in table 5.1, while the dashed line shows the predictions based on the solubility eqn. 5.2.

In the North Atlantic gyre (Fig. 5.11a), an increased Northern boundary temperature results in a larger $\overline{pCO_2}$ anomaly. However, the response is generally small compared to the expected trend given by solubility. Only in the extreme Northern warming case, $++T_N$, does the effect of the solubility become significant such that the modelled $\overline{pCO_2}$ is just over 50% that of the predicted value using the solubility relationship. We find that in the North Atlantic, the dynamical effect of the surface warming on $\overline{pCO_2}$ is of the same order and opposite sign as the solubility effect.

In the case of Southern boundary warming (cyan in Fig. 5.11), the stratification in the Southern Ocean is increased which inhibits the upwelling and mixing of carbon-rich

water to the surface (as in eqn. 3.7). The decrease in upwelling causes a negative effect in the dynamical contribution of Southern Ocean $p\text{CO}_2$ to warming, lowering $\overline{p\text{CO}_2}$. In the case our Southern boundary temperature perturbations, the buoyancy perturbation is confined to the Southern Ocean so there is little change in North Atlantic $\overline{p\text{CO}_2}$ as the Southern boundary warms.

When comparing the relative response of the Northern and Southern high latitude $\overline{p\text{CO}_2}$ to warming at the Northern boundary, shown in Fig. 5.12, we find that as the Northern $\overline{p\text{CO}_2}$ increases, the Southern Ocean $\overline{p\text{CO}_2}$ reduces. The MOC depends on the shared buoyancy window between the North Atlantic sub-polar gyre and Southern Ocean [Gnanadesikan, 1999, Nikurashin and Vallis, 2012]. As the Northern pole warms, the shared buoyancy range is reduced⁴, however, the residual circulation in the Southern Ocean varies with the zonal wind stress (and subsequent eddy response), which remains mostly constant with changes in the boundary temperatures. The system must therefore be closed by diapycnal mixing [Gnanadesikan, 1999]. As the Northern pole warms, the MOC becomes more decoupled from the Southern Ocean. More carbon remains in Northern sub-polar gyre mixed layer and less Northern hemisphere carbon is upwelled in the Southern Ocean, explaining the observed pattern in the $\overline{p\text{CO}_2}$ response. Similarly, more heat is stored in the Northern hemisphere as opposed to the Southern hemisphere. The result is a North-South dipole pattern whereby the polar warming causes a positive saturated carbon anomaly in the North and a negative anomaly in the Southern Ocean (e.g. Fig. 5.15a).

The North-South dipole response in high latitude $\overline{p\text{CO}_2}$ to polar warming is an example of how dynamics can induce changes in the carbon system that compensate each other to a certain degree. However, the overall atmospheric $p\text{CO}_2$ feedback due to natural carbon is not set by the high latitude $p\text{CO}_2$ response (as shown in sections 5.1.2 and 5.1.3), but by global solubility changes. In response to warming, the atmospheric heat anomaly is absorbed throughout the whole upper ocean. As a result, global solubility of carbon is reduced and it is the global reduction in solubility which determines the atmospheric response [Joos et al., 1999, Goodwin and Lenton, 2009]. In terms of natural carbon, we expect that the high latitudes will therefore only play a small role in setting the

⁴The ocean density equation can be written as a linear equation in terms of temperature, such that warming would shift the overlap region between the Northern and Southern density ranges.

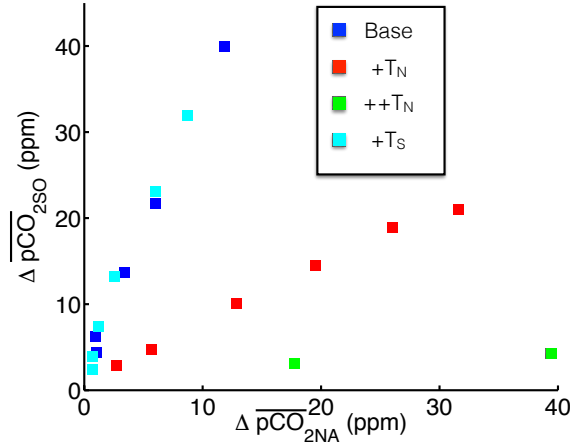


Figure 5.12: Change in Southern Ocean $p\text{CO}_2$ as a function of the change in North Atlantic $p\text{CO}_2$ for various warming scenarios.

atmospheric response. Accompanying the solubility changes, large scale ocean dynamics will adjust to the change in buoyancy forcing. These dynamical changes cause additional anomalies in saturated, regenerated and disequilibrium carbon. We can use the suite of MITgcm warming experiments to understand how the ocean carbon reservoirs respond to the warming-induced dynamical adjustments.

5.2 Carbon compensation and the atmospheric feedback

5.2.1 Carbon reservoirs in warming scenarios

The saturated carbon reservoir has a strong dependency on ocean heat content. However, the disequilibrium carbon reservoir has a more direct dependence on the global overturning circulation than the saturated carbon reservoir. Additionally, the ventilation of regenerated carbon is also determined by the depth and strength of the large scale overturning circulation. Therefore by perturbing the ocean dynamics through polar warming, we can probe the response of the regenerated, saturated and disequilibrium carbon reservoirs. In the next section, we will use surface warming and salinity MITgcm perturbation experiments, as well as simple box models to quantify the extent to which saturated, disequilibrium and regenerated carbon changes compensate each other to determine the resulting atmospheric $p\text{CO}_2$ anomaly discussed above. We observe that different compensations occur over different timescales. To illustrate the relevant timescales, Fig. 5.13 shows the time evolution of the atmospheric $p\text{CO}_2$ anomaly in the $1.2\times T$ base warming experiment. There is a clear separation between two phases of adjustment in the system: an initial increase to the maximum value of $\Delta p\text{CO}_2^a$ (blue) followed by a decline over the

following 2600 years to reach an equilibrium after 3000 years of the warming perturbation (red).

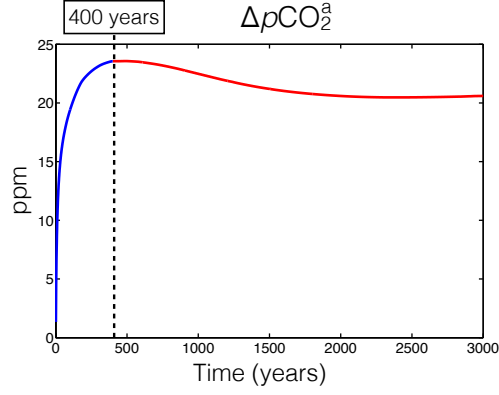


Figure 5.13: Time series of the atmospheric $p\text{CO}_2$ anomaly in the $1.2xT$ base warming experiment. The blue line shows the first 400 years and the red line shows the subsequent 2600 years.

Due to the presence of both a centennial and a millennial timescale adjustment of atmospheric $p\text{CO}_2$, we will analyse the ocean's carbon response at 400 years and at 3000 years.

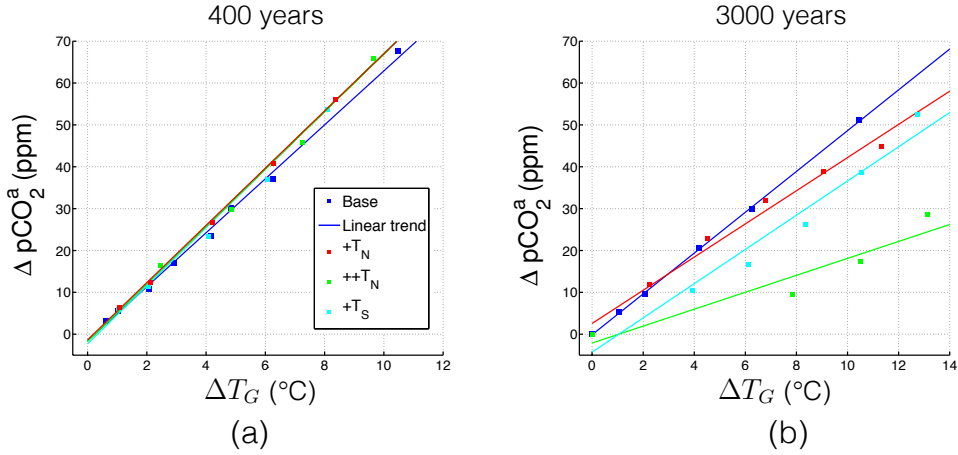


Figure 5.14: Change in atmospheric $p\text{CO}_2$ in MITgcm warming experiments as a function of global mean sea surface temperature anomaly ΔT_G after a) 400 years and b) 3000 years. Squares indicate model data from warming experiments according to the colours listed in table 5.1. The solid lines show the corresponding linear fits.

The response of $p\text{CO}_2^a$ to global mean temperature anomalies at 400 and 3000 years is shown in Figs. 5.14a and 5.14b, respectively. Figure 5.14a shows that $\Delta p\text{CO}_2^a$ follows the same linear trend regardless of the boundary conditions after 400 years. The result

was also tested with a domain 40° longitude in width, with a mean MOC of 6.2 Sv and a AABW cell of 2.2 Sv, and the same trend was found. The trend in $\Delta p\text{CO}_2^a$ implies the polar warming-induced dynamical changes have little impact on the net change in ocean carbon after 400 years. However, after 3000 years, $\Delta p\text{CO}_2^a$ has decreased and the global mean sea surface temperature anomaly has increased compared to the year 400, in all warming experiments as shown in Fig. 5.14b. The reduction in $\Delta p\text{CO}_2^a$ is stronger in cases of both Southern Ocean and North Atlantic warming. Therefore on timescales longer than a few hundred years, dynamical changes appear to have a significant effect on $\Delta p\text{CO}_2^a$.

To further investigate this result, a set of temperature perturbation experiments are run whereby the base warming perturbation is imposed, and the Eulerian and eddy velocity fields are prescribed to the control field, such that the circulation is fixed (although temperature and salinity and therefore density is allowed to vary). By doing so, we can isolate the contribution of dynamics, particularly advection, on the carbon system. The experiment shows that over 400 years, the dynamical contribution is less than 5 ppm at 10° of warming (Fig. 5.16a). The results can be explained by the differences in saturated, disequilibrium and regenerated carbon between the fixed dynamics and freely evolving experiments (Fig. 5.15). An unperturbed overturning means a higher ocean heat uptake, causing a loss in saturated carbon (region 1 in Fig. 5.15a), except in the Southern Ocean where the control residual circulation is stronger such that region 2 is cooler in the fixed dynamics experiment. The stronger residual circulation exports deep ocean carbon from the Southern channel into intermediate waters more efficiently resulting in a positive disequilibrium carbon anomaly (region 3 in Fig. 5.15b). The saturated and disequilibrium carbon anomalies mostly compensate each other over timescales of centuries to millennia (the picture of Fig. 5.15a-b is therefore roughly constant in time).

The most significant dynamical contribution to the ocean carbon distribution in terms of the atmospheric CO_2 response, over all timescales, is the isolation of regenerated carbon in the Northern Hemisphere. The regenerated carbon is isolated because the upper overturning circulation cell shallows and weakens in warming scenarios (region 5 in Fig. 5.15c) and because regenerated carbon is accumulated in the weakening bottom cell. At 400 years, it is partly balanced by changes in Southern hemisphere biology (region 4) but over time, the carbon anomaly in both region 5 and in the bottom cell grows (see Fig.

5.15d). After 3000 years, the isolation of regenerated carbon represents around 20% of the total net change in $p\text{CO}_2^a$ in the base warming scenario.

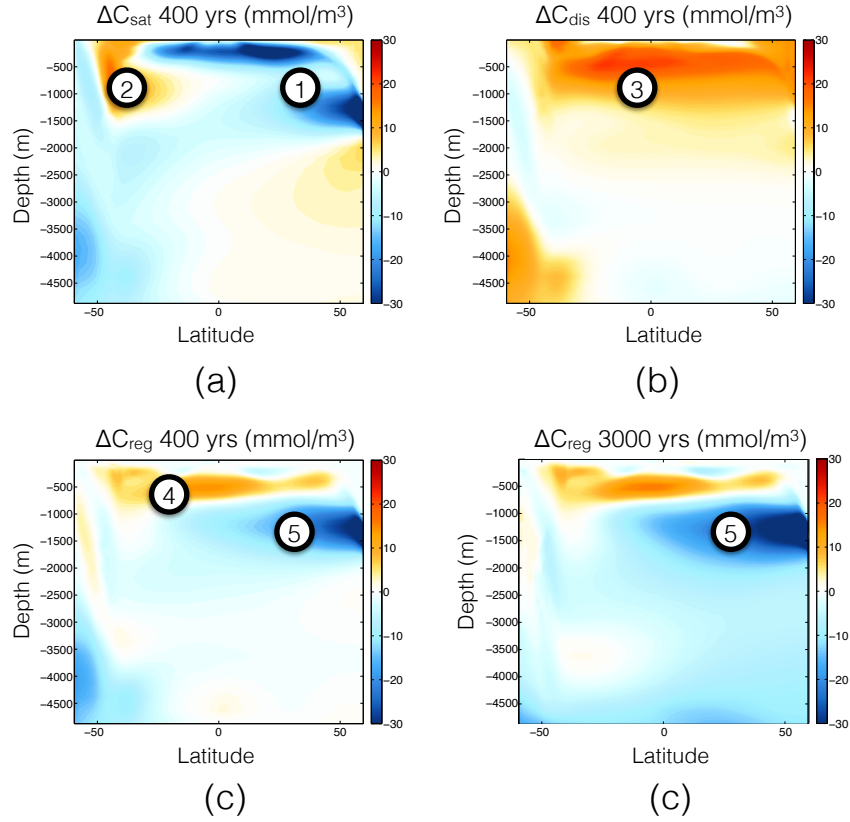


Figure 5.15: Zonal mean difference in a) saturated, b) disequilibrium and c) regenerated carbon after 400 years and d) regenerated carbon after 3000 years between fixed dynamics and freely evolving MITgcm warming experiment under the base warming scenario.

We can compare the dynamical contribution towards $p\text{CO}_2^a$ to that of surface biology. By fixing the global ocean phosphate field we ensure that the nutrient limitation on surface biology is constant. The mixed layer biological activity in the fixed phosphate experiments is constant to within 1% from the control mean. In warming scenarios, surface ocean stratification is increased so the nutrient supply to the surface is reduced, causing a reduction in surface biological productivity and regenerated carbon. In the fixed phosphate scenarios, surface biology is kept at elevated levels such that the loss of ocean regenerated carbon is reduced.

The $p\text{CO}_2^a$ changes in the fixed dynamics and biology experiments are shown in Fig. 5.16a and Fig. 5.16 b at 400 and at 3000 years, respectively. The relative contributions of each of the components (dynamics, solubility and biology) are shown in Fig. 5.16c-d.

The results shows that after 400 years, the contribution of dynamics to the $p\text{CO}_2^a$ anomaly is small (less than 10% of the total $p\text{CO}_2^a$ anomaly). While the $p\text{CO}_2^a$ anomaly is largely due to the solubility, changes in surface biology contribute roughly 25% towards the total. We can calculate the theoretical trend in $p\text{CO}_2^a$ using the framework from Goodwin et al. [2008] (dashed lines in Figs. 5.16a and 5.16b) and compare the theoretical trend with the results from the fixed nutrient experiments, which match the prediction⁵.

The results indicate that after 400 years of warming, the $p\text{CO}_2^a$ response is dominated by solubility, which accounts for roughly 75% of the total, with the other 25% being due to the biological productivity changes, consistent with older studies such as Joos et al. [1999]. Over this timescale, the effects of dynamical changes cause compensating responses in disequilibrium and saturated carbon such that the dynamics have a small effect on atmospheric $p\text{CO}_2$. On millennial timescales, however, the strength of the atmospheric $p\text{CO}_2^a$ feedback due to solubility and biology remain roughly constant, but the circulation-induced isolation of regenerated carbon increases and causes an overall reduction in the $p\text{CO}_2^a$ anomaly.

Figure 5.14b shows that as the Northern and Southern boundaries warm more, the dynamical contribution increases in strength after millennia and counteracts the effects of biology and solubility. The most relevant dynamical change induced by the polar warming for the compensation of the carbon reservoirs over millennia is the strength and depth of the meridional overturning circulation and the bottom cell, as shown in Fig. 5.10. We can therefore further explore the mechanisms of the millennial carbon compensation by perturbing the Northern and Southern boundary salinities. By doing so, we alter the high latitude buoyancy which will cause similar changes to the large scale circulation as the warming experiments, but without the solubility effects associated with warming. Moreover, this approach isolates contributions from the North Atlantic and from the Southern ocean, without the mid-latitude buoyancy forcing imposed in the warming experiments. The next section discusses the results of the salinity experiments and how we can use them to understand compensation of carbon reservoirs over millennial timescales due to changes in the large scale overturning.

⁵We use the equation: $p\text{CO}_2^a = p\text{CO}_{2i}^a \exp\left(-\frac{V}{I_B} \frac{\partial C_{sat}}{\partial T} \Delta T_O\right)$ where $p\text{CO}_{2i} = 302$ ppm is the initial $p\text{CO}_2$, V is the ocean volume, I_B is the buffered carbon inventory and ΔT_O is the change in global ocean mean temperature, averaged over V , as in section 1.3.2. We used values of $V/I_B = 0.98$ and $\partial C_{sat}/\partial T = -0.1 \text{ gm } ^{-3}\text{C}^{-1}$ [Goodwin et al., 2008].

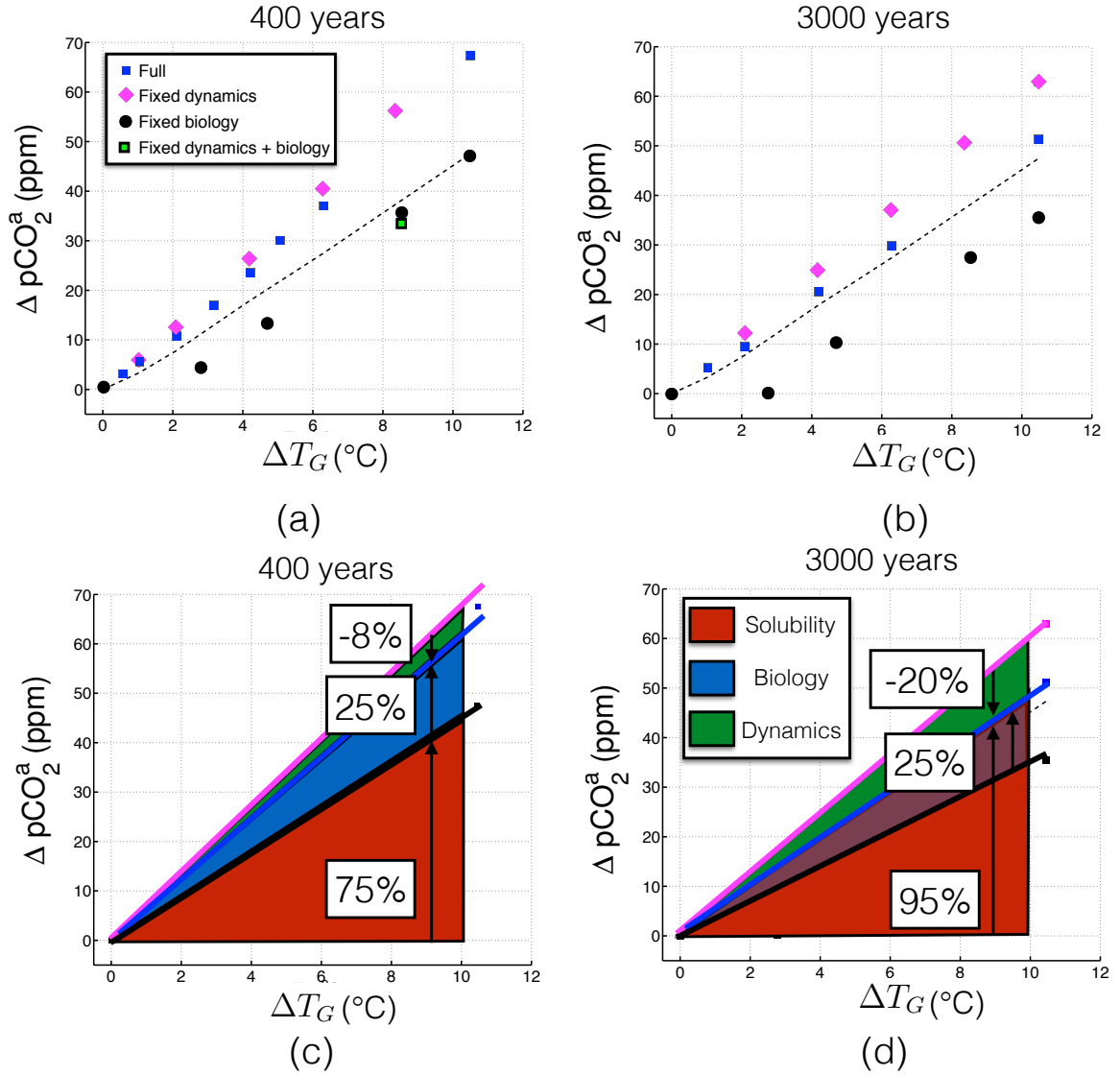


Figure 5.16: Change in atmospheric $p\text{CO}_2$ in MITgcm as a function of global mean sea surface temperature anomaly ΔT_G after a) 400 years and b) 3000 years of experiment. Blue squares represent the base warming scenario, magenta shows fixed dynamics experiments and black data shows fixed phosphate runs. The dashed black line shows the theoretical solubility-driven trend based on Goodwin et al. [2008]. Panels c)-d) show the same as a)-b) but with solid lines which linear fits to MITgcm data and coloured triangles overlain to show the respective contributions of solubility, dynamics and biology to setting the overall $p\text{CO}_2^a$. The percentages shown are approximate ($\pm 10\%$).

5.2.2 Isolating the effect of the MOC

North Atlantic buoyancy perturbation experiments have been performed in the past using a wide variety of models. The most notable dynamical change relevant to the carbon cycle is the decrease in the pole-to-pole meridional overturning circulation (MOC) [Stommel, 1961, Stocker and Wright, 1991, Rahmstorf, 1996, Ganopolski and Rahmstorf, 2001,

Saenko et al., 2003, Wolfe and Cessi, 2011]. In the perturbation experiments presented here, we decrease the Northern boundary salinity restoring S_N from 34 psu to 32 psu and perform model integrations for 5000 years. Over the range studied, the strength of the MOC decreases linearly from 4.2 Sv in the control run with $S_N = 34$ psu to an effective shutdown at $S_N = 32$ psu (Fig. 5.17a). The depth at which the maximum MOC strength occurs in the North Atlantic shows an overall shallowing from 650 m to 400 m depth (Fig. 5.17b).

Since only the northern boundary is perturbed, the bottom circulation originating at the Southern boundary remains roughly constant with a strength of 1.2 – 1.5 Sv (Fig. 5.17c). Therefore the experiments effectively isolate the role of the MOC on atmospheric carbon, without a large influence from the bottom water cell. Since the salinity perturbations imposed aim to reduce surface buoyancy, isopycnals are depressed leading to a deepening of the pycnocline (Fig. 5.17d).

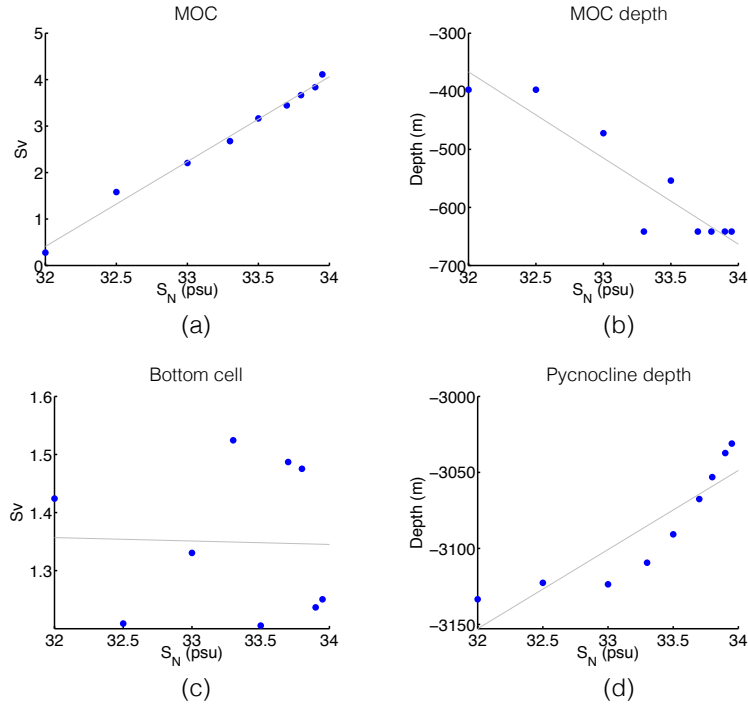


Figure 5.17: Diagnostics in the MITgcm Northern salinity perturbation experiments as a function of Northern boundary salinity restoring S_N : a) Strength of the MOC, b) Depth of the MOC maximum, c) Bottom overturning cell strength and d) Pycnocline depth.

In the control run (with $S_N = 34$ psu), the MOC consists of a pole-to-pole overturning circulation extending from the Northern boundary to the Southern channel (Fig. 5.18a).

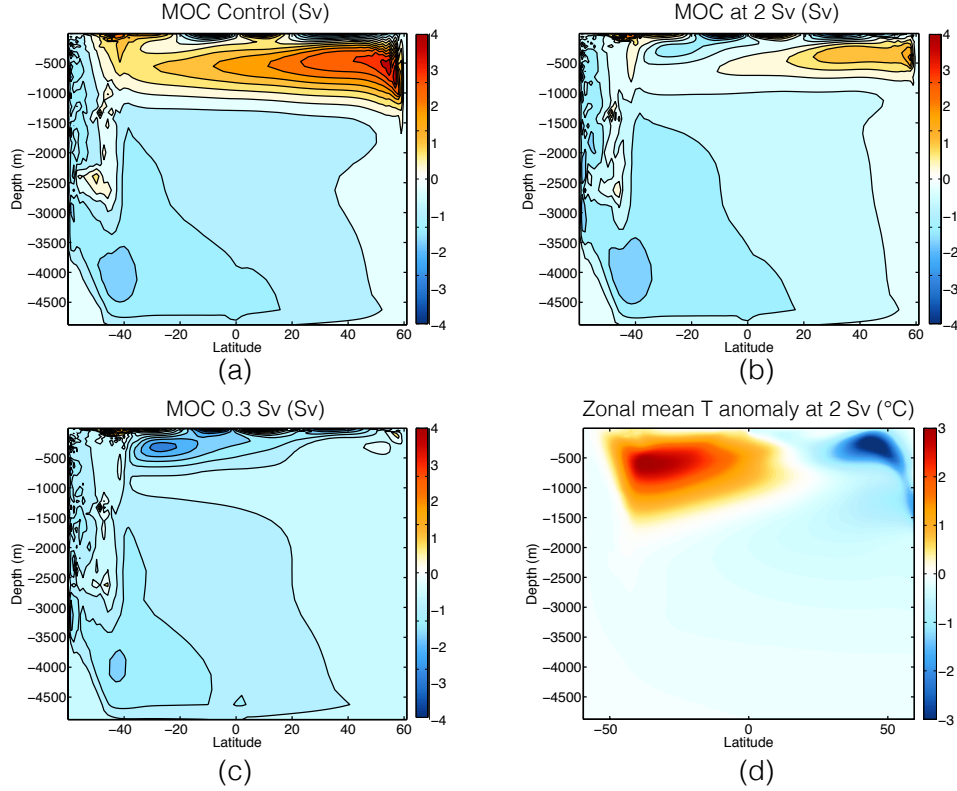


Figure 5.18: Meridional overturning stream function (in Sv) after 5000 years for a) control run strength of 4.2 Sv, b) 2 Sv ($S_N = 33$ psu) and c) 0.1 Sv ($S_N = 32$ psu); and d) the zonal mean temperature anomaly after 5000 years of the $S_N = 33$ psu MITgcm Northern salinity perturbation experiment, relative to the control run (in °C).

A freshening of the Northern boundary, induced by a weaker salinity restoring value, reduces the range of isopycnals shared between the Southern channel and the Northern sub-polar gyre, resulting in a disconnect between the northern and southern high latitude regions. For a salinity forcing of $S_N = 33$ psu, Fig. 5.18b shows the zonally averaged overturning streamfunction as a function of latitude and depth, corresponding to an MOC strength of 2 Sv at the end of the experiment, and illustrates a partial disconnect while still preserving the diffusive MOC in the Northern Hemisphere (see Fig. 5.3a in section 5.1.1). Figure 5.18c shows a complete shutdown of the MOC. As the density in the sub-polar gyre increases, the intermediate water formed north of the Southern Ocean (Antarctic Intermediate Water) gradually becomes denser than water in the North Atlantic and a shallower Southern Hemisphere overturning mode is favoured [Manabe and Stouffer, 1999, Ganopolski and Rahmstorf, 2001, Gregory et al., 2003, Saenko et al., 2003].

The carbon cycle response to a salinity-induced North Atlantic buoyancy change is complex due to the various processes and associated timescales responsible for different

carbon reservoirs. A decrease in salinity forcing leads to a weakening of the MOC and several changes in dynamical and biogeochemical processes:

1. The increase in North Atlantic buoyancy causes a disconnect between the Northern hemisphere and Southern hemisphere MOC, as described above. As a result, less cold deep water from the Northern hemisphere is transported to the Southern Ocean. Instead, warmer water from tropical origin is circulated within the Southern Hemisphere, causing warming in the upper 1500m of the Southern hemisphere (see Fig. 5.18). The warming causes a reduction in the solubility of carbon, resulting in an initial decrease in ocean saturated carbon C_{sat} (see region 1 in Fig. 5.19 and the timeseries in Fig. 5.20).
2. The reduction of the upwelling rate in the ocean interior reduces the ventilation of deep carbon-rich waters, increasing saturated and preformed carbon concentrations (region 2 in Fig. 5.19). The reduction in northwards heat transport and convection in the North Atlantic causes a reduction in local ocean heat uptake (see Fig. 5.18d), increasing the solubility of carbon and therefore C_{sat} (region 3 in Fig. 5.19) [Xie and Vallis, 2012].
3. The reduction in surface biological productivity reduces the global production of biological carbon and C_{reg} (region 4 in Fig. 5.19). This is especially prevalent in the Southern Hemisphere due to the disconnect between the Southern and Northern hemisphere circulation. At the same time, the shallowing and weakening of the MOC isolates the biological carbon at depth in the Northern Hemisphere (region 5 in Fig. 5.19). Overall, the reduction in biological carbon due to the reduced biological productivity is stronger and causes a net decrease in C_{reg} [Schmittner, 2005, Schmittner et al., 2007, Zickfeld et al., 2008]. The decrease in C_{reg} eventually causes the upwelling water to contain less biological carbon, which in turn results in a decrease in C_{dis} (e.g. region 6 in Fig. 5.19).

The processes affecting the carbon reservoirs described above are listed in order of timescales: such that process 1 acts on decadal to centennial timescales, process 2 acts on centennial timescales and processes 2 and 3 act on millennial timescales. The time evolution of each of the reservoirs in the salinity perturbation experiment with $S_N = 33$ psu, corresponding to a 2.1 Sv MOC is shown in Fig. 5.20a. Initially, there is a reduction in C_{sat} anomalies, as described in process 1 above, followed by a long term restoring toward

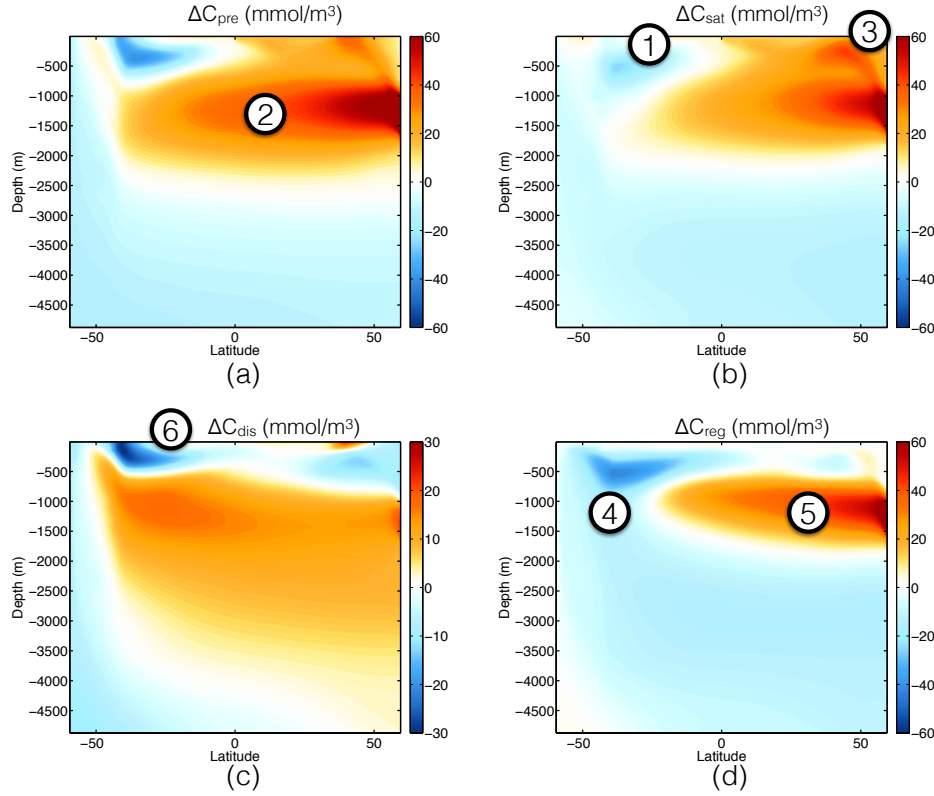


Figure 5.19: Carbon cycle components anomalies in the $S_N = 33$ psu MITgcm Northern salinity perturbation experiment after 5000 years relative to the control run, in mmol/m^3 . The final MOC in the forced experiment is 2.1 Sv. The panels show zonally averaged anomalies of: a) C_{pre} , b) C_{sat} , c) C_{dis} and d) C_{reg} , as function of depth and latitude.

zero due to process 2. The initial increase in the disequilibrium carbon component, C_{dis} , is due to the reduction of the circulation and isolation of disequilibrium carbon described in process 2. However, after 1500 years the global reduction in the biological carbon reservoir causes a decline in C_{dis} as upwelling waters hold less excess biological carbon. The decline in the biological carbon can be seen by the long term decrease in C_{reg} anomalies. Figure 5.20 also shows that while the *total* carbon stored in each of the carbon reservoirs is vastly different (see section 1.2.1), the anomalies in each carbon reservoir are comparable to each other in magnitude.

The interaction of timescales results in an oscillatory-like behaviour of the atmospheric $p\text{CO}_2$ (Fig. 5.20b). An initial increase in $p\text{CO}_2^a$ due to ocean loss of C_{sat} is followed by a decline in $p\text{CO}_2^a$ due to the ocean increase in C_{dis} . Finally, there is a long term decrease in C_{reg} . The net result over millennia is a partial balance between C_{pre} and C_{reg} , of which the loss in C_{reg} is larger, so the $p\text{CO}_2^a$ anomaly is positive [Schmittner, 2005, Schmittner et al.,

2007, Zickfeld et al., 2008]. Overall, the atmospheric $p\text{CO}_2$ anomaly is driven by saturated carbon on timescales less than 1000 years, but afterwards the anomaly is driven by the regenerated carbon. However, since the relative changes in C_{reg} and C_{dis} are comparable, atmospheric $p\text{CO}_2$ does not increase by more than 5ppm after several thousands of years. Such a change is less than seen in other modelling studies with global GCMs such as Schmittner et al. [2007], Zickfeld et al. [2008] the reasons for which will be explored later in section 5.2.4.

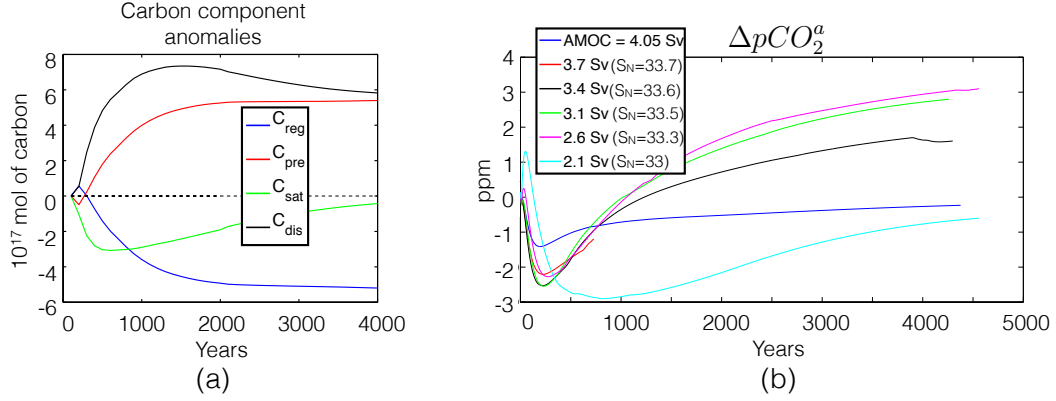


Figure 5.20: Carbon cycle response to salinity perturbations: a) ocean carbon cycle component anomalies in the $S_N = 33$ psu experiment (with an overturning of 2.1 Sv) relative to the control run and b) atmospheric $p\text{CO}_2$ as a function of time for a range of MOC strengths in the MITgcm Northern salinity perturbation experiments.

5.2.3 Isolating the effect of the bottom cell

Similar to the North Atlantic salinity perturbation experiments, we run a set of experiments whereby the Southern boundary salinity S_S is reduced from 34 psu to 33.2 psu in increments of 0.2 psu. The most significant change is the reduction in the bottom overturning cell from a control value of 1.3 Sv to an effective shutdown at 0.2 Sv (Fig. 5.21a). The change in Southern boundary density also has an effect on the MOC, which increases as the bottom cell weakens (Fig. 5.21b). The effect is sometimes referred to as the bipolar seesaw [Stocker et al., 1992, Seidov et al., 2001, Brix and Gerdes, 2003, Swingedouw et al., 2009], although it is model-dependent [Seidov et al., 2005]. As the Southern Ocean becomes less dense, the range of isopycnals shared with the North Atlantic sub-polar gyre increases, leading to an increase in the pole-to-pole overturning [Gnanadesikan, 1999, Wolfe and Cessi, 2011], shown in Fig. 5.21b. The positive surface buoyancy anomaly spreads throughout the basin and depresses isopycnals, leading to a deepening of the pycnocline (Fig. 5.21c), as with the Northern boundary perturbation experiments.

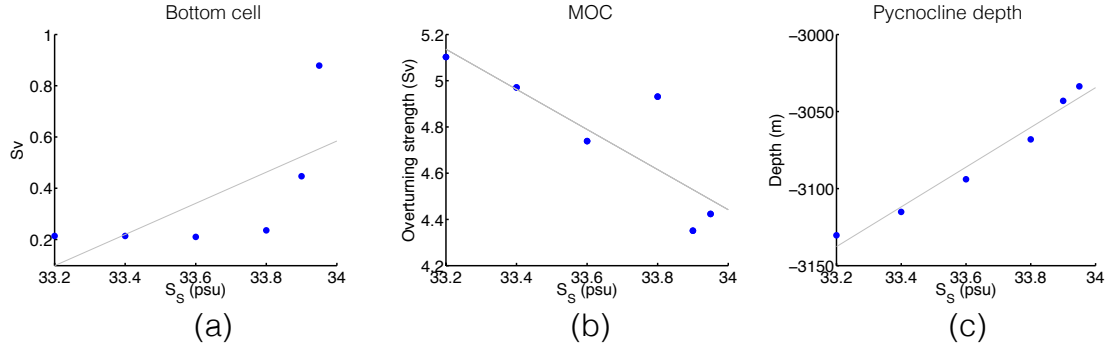


Figure 5.21: Diagnostics in the MITgcm Southern salinity perturbation experiments as a function of Southern boundary salinity restoring S_S : a) Strength of the bottom cell (Sv), b) Strength of the MOC (Sv) and c) Pycnocline depth (m).

The effects of the dynamical changes induced by the Southern Ocean restoring anomaly can be summarised by considering their effect on the carbon cycle components:

1. The reduction in deep water formation at the Southern boundary results in the normally oversaturated subducted waters to be more equilibrated with the atmosphere, leading to a reduction in C_{dis} (Fig. 5.22c).
2. Since most of the nutrients and organic matter are remineralized in the upper 2000m of the ocean, the MOC is mostly responsible for re-supplying these nutrients to the surface to fuel biological productivity. On the other hand, the bottom water cell is mostly responsible for flushing biological carbon back into the surface ocean and atmosphere [Marinov et al., 2006, 2008b]. Therefore, as the bottom water overturning cell weakens, it accumulates biological carbon (Fig. 5.22d, note the difference in scales).
3. The remineralization of biological carbon also increases alkalinity. Therefore, as the bottom cell circulation slows down, alkalinity concentrations build up (Fig. 5.22e) at the expense of alkalinity in the upper cell. Since higher alkalinity waters can hold more carbon (due to an increased buffering capacity), the pattern of alkalinity anomalies dominate the changes in the saturated carbon reservoir (Fig. 5.22b).

The timescales associated with the overturning and mixing of the bottom water cell are hundreds to thousands of years, so the listed processes affecting the carbon reservoirs all act over similar millennial timescales. Figure 5.23a shows the time evolution of C_{pre} ,

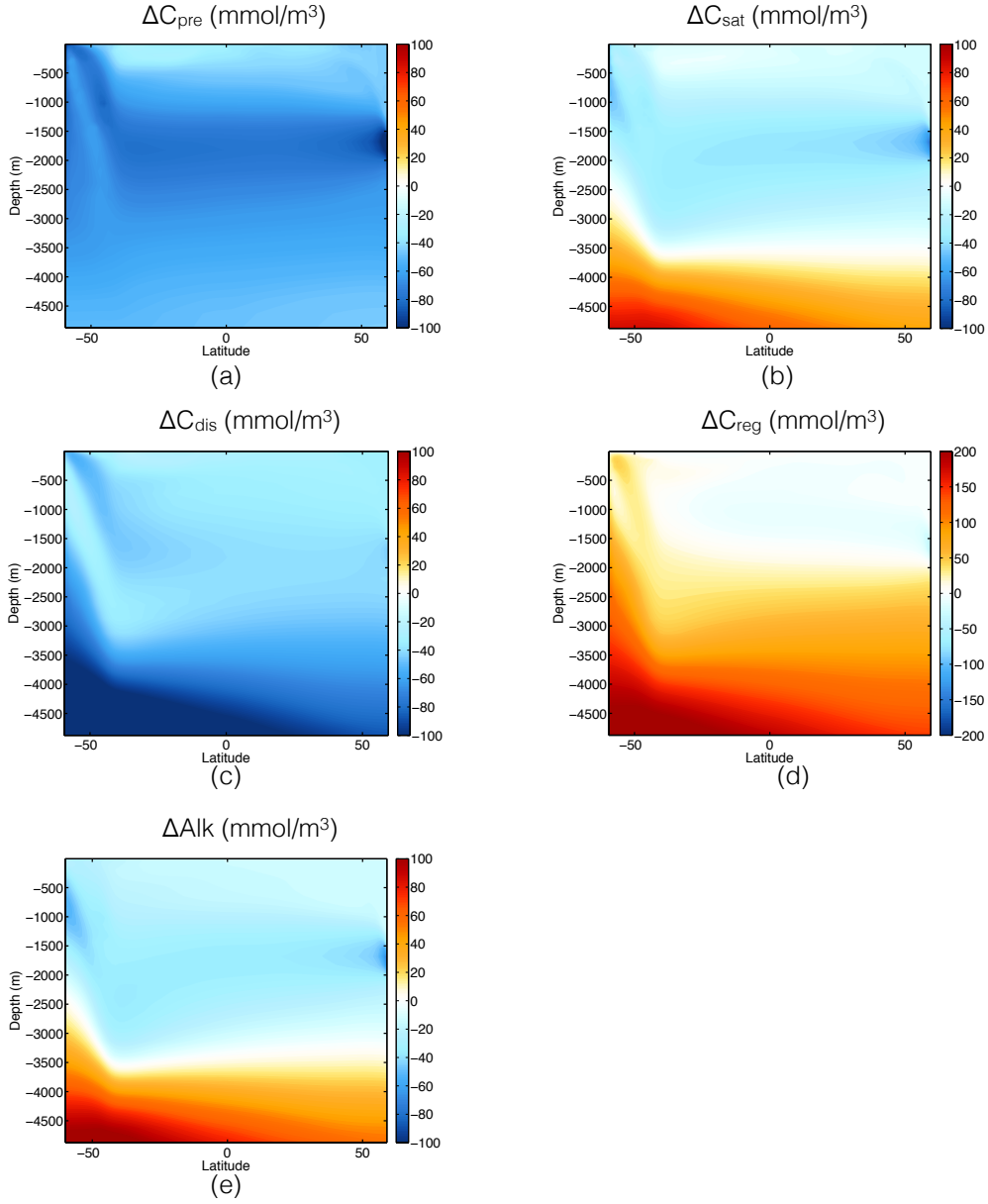


Figure 5.22: Carbon cycle component anomalies for the $S_S = 33.6$ psu MITgcm Southern salinity perturbation experiment, corresponding to a bottom cell strength of 0.2 Sv (after 5000 years) relative to the control run, in mmol/m³: a) C_{pre} , b) C_{sat} , c) C_{dis} , d) C_{reg} and e) alkalinity.

C_{sat} , C_{dis} and C_{reg} anomalies for a salinity forcing of $S_S = 33.6$ psu, corresponding to a bottom cell strength of 0.2 Sv at the end of the experiment. The preformed carbon anomaly is dominated by the disequilibrium component. However, the preformed carbon anomaly is smaller than the regenerated carbon anomaly over all timescales. The net result is therefore an increase in ocean carbon content and a decrease in atmospheric $p\text{CO}_2$ as shown in Fig. 5.23b. The atmospheric $p\text{CO}_2$ anomalies have a larger amplitude than

$p\text{CO}_2^a$ anomalies arising from a weakening of the MOC. $p\text{CO}_2^a$ anomalies due a weakening of the bottom cell reach -45ppm, revealing a stronger climate-carbon feedback than due to reductions in the North Atlantic MOC. The stronger atmospheric CO_2 feedback due to the bottom cell is firstly due to the amount of regenerated carbon the bottom cell contains. There is therefore less compensation between the regenerated and preformed reservoirs than in the case of MOC changes. Secondly, the upwelling of bottom water has a fundamentally different influence on surface biological productivity. The majority of nutrients are remineralized in the upper cell (due to the exponential decay of the vertical nutrient flux with depth [Martin et al., 1990]) which are re-circulated and supplied to the upper ocean North of the Southern Ocean. However, Antarctic Deep Water circulates mostly south of the Antarctic circumpolar current, so the influence on global biology is small. Yet, the DIC concentration of deep Antarctic water is high, so its potential to influence atmospheric $p\text{CO}_2$ is high (e.g. Marinov et al. [2006, 2008b]).

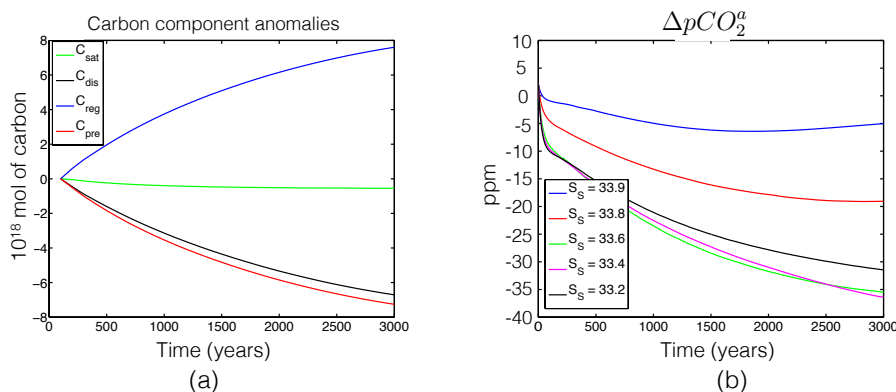


Figure 5.23: First 3000 years of the time series of the carbon cycle response to salinity perturbations: a) ocean carbon cycle components anomalies for the $S_S = 33.6$ psu experiment (bottom cell strength of 0.2 Sv) relative to the control and b) atmospheric $p\text{CO}_2$ for a range of Southern boundary salinities.

5.2.4 Predicting carbon compensation using box models

The MITgcm experiments show that the magnitude of the atmospheric $p\text{CO}_2$ change due to a high latitude buoyancy perturbation is set by the partial balance in the gain and loss of C_{reg} and C_{pre} . We can predict how atmospheric carbon varies with the strength of the overturning by constructing simple box models for each carbon reservoir. Using the box models, we can therefore predict the compensation between the carbon reservoirs. We will use a box model for preformed carbon and for nutrients which are simplified versions of

those used in Toggweiler [1999], Toggweiler et al. [2003a] and Bacastow [1996] (see also section 2.4.2).

For preformed carbon, the model consists of two ocean boxes: a surface box of thickness h with carbon concentration C_S (mol/m³) and pCO_2^S (ppm), along with a deep box of thickness H with carbon concentration C_D . The layers interact through an overturning circulation of strength Ψ and mixing of strength f . The mixing, f , parameterises both diapycnal and isopycnal vertical mixing processes, such as convective mixing in the Southern Ocean, equatorial upwelling, eddy stirring or diapycnal upwelling in the ocean interior. The surface ocean exchanges carbon with a well-mixed atmospheric box with a partial pressure of CO₂ given by pCO_2^a through a flux $F_{air-sea}$, as shown schematically in Fig. 5.24a, where $F_{air-sea} = k(pCO_2^a - pCO_2^S)$ and k is the gas exchange parameter [Wanninkhof, 1992].

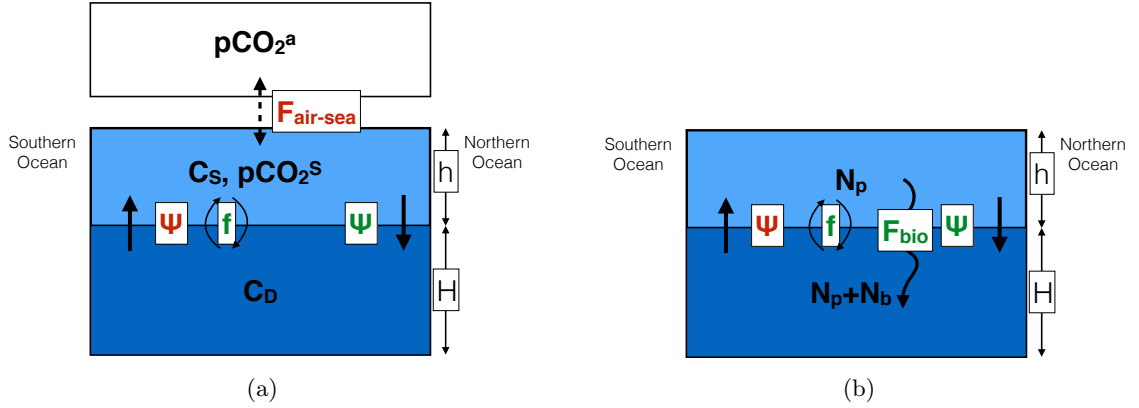


Figure 5.24: Transport box models for a) preformed carbon and b) nutrients, derived from Toggweiler [1999], Toggweiler et al. [2003a]. The preformed box models consists of a surface box of thickness h with carbon concentration C_S (mol/m³) and pCO_2^S (ppm), along with a deep box of thickness H with carbon concentration C_D and an atmosphere with a partial pressure of CO₂ given by pCO_2^a . The nutrient model has a surface ocean box with nutrient concentration N_p and a deep box with concentration equal to $N_p + N_b$. Tracers are exchanged between the box through an overturning of strength Ψ , vertical mixing of strength f , an air-sea CO₂ flux $F_{air-sea}$ and biological flux F_{bio} .

At equilibrium, carbon transport through advection and mixing in the surface box must be balanced by an air-sea flux such that

$$\Psi C_D - \Psi C_S + f(C_D - C_S) = k(pCO_2^a - pCO_2^S), \quad (5.4)$$

where we assume that $C_D > C_S$ and that the atmosphere has other sources and sinks of carbon which are constant (e.g., the land surface processes). Since we are looking for

perturbations in $p\text{CO}_2^a$ over the pre-industrial state, namely $\Delta p\text{CO}_2^a$, we will differentiate eqn. 5.4. As will become apparent later on, the response of the carbon cycle depends on the relative magnitudes of the overturning and mixing strength, i.e. Ψ/f . Therefore we choose at this stage to differentiate with respect to Ψ/f and express the equation in the form:

$$\Delta p\text{CO}_2^a = \frac{f}{k} [C_D - C_S] \Delta (\Psi/f), \quad (5.5)$$

where $\Delta (\Psi/f) = (\Psi/f - \langle \Psi/f \rangle)$ and “ $\langle \dots \rangle$ ” indicates the pre-industrial mean. We assume that the ocean carbon concentrations C_D and C_S remain the same, as well as the global surface ocean $p\text{CO}_2^S$.⁶ The sensitivity of the atmospheric $p\text{CO}_2$ anomaly to changes in Ψ will therefore depend on the difference between deep and surface carbon concentrations, the strength of the mixing f , and the air-sea gas exchange k . For natural carbon, with C_D larger than C_S , the overturning and mixing effectively erode gradients between the two reservoirs, transferring carbon from the deep into the surface box. Therefore, in this simple formulation, an increase in overturning circulation would lead to an increase in atmospheric $p\text{CO}_2$.

A two-box model can be used to understand the response of nutrients to changes in the overturning Ψ and mixing f . The surface and deep box can exchange nutrients via Ψ and f as above but also through a biological flux F_{bio} , as shown in Fig. 5.24b. The surface ocean box has a nutrient concentration N_p while the deep box has a concentration equal to $N_p + N_b$, where N_b is the biological component and N_p the preformed component. The biological flux depends on the amount of nutrients upwelled, $\Psi(N_p + N_b)$, and surface limitation of light and nutrients γ , which takes a value between 0 and 1 [Tilman, 1977]. We choose to make the surface biology only dependent on the nutrients upwelled via the overturning circulation Ψ and not the mixing f for the following reasons:

- In our formulation, we separate the effects of upper and lower overturning cells by including the bottom overturning cell as part of the vertical mixing f .
- The downward flux of nutrients below the mixed layer decreases with depth [Martin et al., 1987], therefore the majority of nutrient are remineralized in the upper

⁶We can make these assumptions because the global ocean contains 50 times more carbon than the atmosphere, so a small change in ocean carbon concentrations can still result in a large change in atmospheric $p\text{CO}_2$.

overturning cell and returned to the surface. The upper cell also has a faster overturning timescale than the bottom cell so is therefore more efficient at ventilating the nutrient-rich deep ocean.

- Marinov et al. [2006] show that the regions South of Antarctic Circumpolar Current (where Antarctic Bottom Water is formed) are important for the air-sea balance of CO_2 , while the regions North of the ACC (where Intermediate Water is formed as part of the MOC) are important for global export production of carbon. Therefore the MOC is more important for the supply of surface nutrients, whereas the bottom overturning cell largely outgasses excess biological carbon.

Conservation of nutrients in the surface ocean therefore requires that:

$$\Psi [N_p + N_b] + f N_b = \gamma \Psi [N_p + N_b] + \Psi N_p. \quad (5.6)$$

The biological nutrient component, N_b , refers to the nutrients which are stripped from the surface layer and exported to the deep ocean via F_{bio} . Therefore only the preformed component N_p remains at the surface. Similarly, the difference in concentration between the surface and bottom layer is due to biology (without biology, the nutrients would be evenly mixed). Since the concentration of preformed nutrients N_p is generally an order of magnitude larger than the biological nutrient concentration [Volk and Hoffert, 1985, Ito and Follows, 2005], the biological flux depends mostly on the overturning strength for nutrient supply whereas f is responsible for mixing excess nutrients between the surface and deep boxes. Using the definition for the efficiency of the biological pump N^* [Ito and Follows, 2005], we obtain $N^* = \frac{N_b}{N_p + N_b}$ (the total average nutrient concentration is taken as $N_p + N_b$ since H is much larger than h). Using eqn. 5.6, we obtain an expression for N^* given by:

$$N^* = \frac{\Psi \gamma}{\Psi + f}. \quad (5.7)$$

Ito and Follows [2005] derive an expression for the variation in atmospheric $p\text{CO}_2$ with the efficiency of the biological pump:

$$\frac{\partial p\text{CO}_2^a}{\partial N^*} = -\frac{R r_{C:N} [N_p + N_b] p\text{CO}_2^a}{C_D}, \quad (5.8)$$

where R is the global mean Revelle Buffer Factor and $r_{C:N}$ is the Redfield ratio between carbon and nutrients. The perturbation in atmospheric $p\text{CO}_2^a$ as function of f/Ψ is given by:

$$\Delta p\text{CO}_2^a = -\frac{R r_{C:N} [N_p + N_b] p\text{CO}_2^a}{C_D} \gamma \left[\frac{1}{f/\Psi + 1} - \left\langle \frac{1}{f/\Psi + 1} \right\rangle \right]. \quad (5.9)$$

f/Ψ is of order 1 Toggweiler [1999], Toggweiler et al. [2003a], so we cannot linearize the expression for N^* . If the global light limitation factor γ increases, more carbon is sequestered into the ocean and $p\text{CO}_2^a$ goes down so in this case $\Delta p\text{CO}_2^a$ will be negative. On the other hand, if the mixing f increases, more excess biological carbon is mixed from the deep into the surface ocean and $p\text{CO}_2^a$ is increased so in this case $\Delta p\text{CO}_2^a$ will be positive. As the overturning strength Ψ increases, more nutrients are cycled through the surface leading to an increase in biological export and a drawdown of $p\text{CO}_2^a$.

Combining equations 5.5 and 5.9, the net change in atmospheric $p\text{CO}_2^a$ is:

$$\Delta p\text{CO}_2^a = \frac{f}{k} [C_D - C_S] \Delta (\Psi/f) - \frac{Rr_{C:N} [N_p + N_b] p\text{CO}_2^a}{C_D} \gamma \Delta \left(\frac{1}{f/\Psi + 1} \right). \quad (5.10)$$

Averaged values of k , C_D , C_S , R , N_b and N_p can be obtained from models. The values of f and γ are less certain. γ is constrained to be less than 1 (biology cannot be more than 100% efficient). Therefore using eqn. 5.7, we obtain the following constraint

$$N^* \left(1 + \frac{f}{\Psi} \right) < 1. \quad (5.11)$$

For the MITgcm control run, with an estimated $N^* = 0.14$, f/Ψ has to be less than 6. Since γ is usually on the order 0.3 – 0.4 [Ito and Follows, 2005], we calculate the response of $p\text{CO}_2^a$ for a range of values of f/Ψ . The maximum value is given by the constraint 5.11 and the minimum value is set by $N^* = 0.14$, such that γ must be at least 0.14. The most likely value of f/Ψ is therefore 2 – 3, given the parameter values for N^* and γ stated above, similar to the range of Toggweiler [1999] who suggest f/Ψ should be in the range of 2 – 4.

For all possible control values of f/Ψ , we can calculate the total change in $p\text{CO}_2^a$ due to a change in the MOC and both the preformed and biological reservoirs. Figure 5.25 shows the change in $p\text{CO}_2^a$ as a function of the ratio f/Ψ , calculated using eqn. 5.10 for different values of the parameter γ and assuming that f is constant and while Ψ is varying. For all salinity perturbation experiments, we find the constraint on $f/\Psi < 6$ remains relatively constant. If γ and f are constant, as the MOC weakens (and f/Ψ increases), the system will reach a new equilibrium which will lie on a black line of constant γ . For example, consider a control value of f/Ψ indicated by the star in Fig 5.25a. As the MOC weakens, the new equilibrium will lie on the line indicated by the blue arrow (in the case that γ and f are constant).

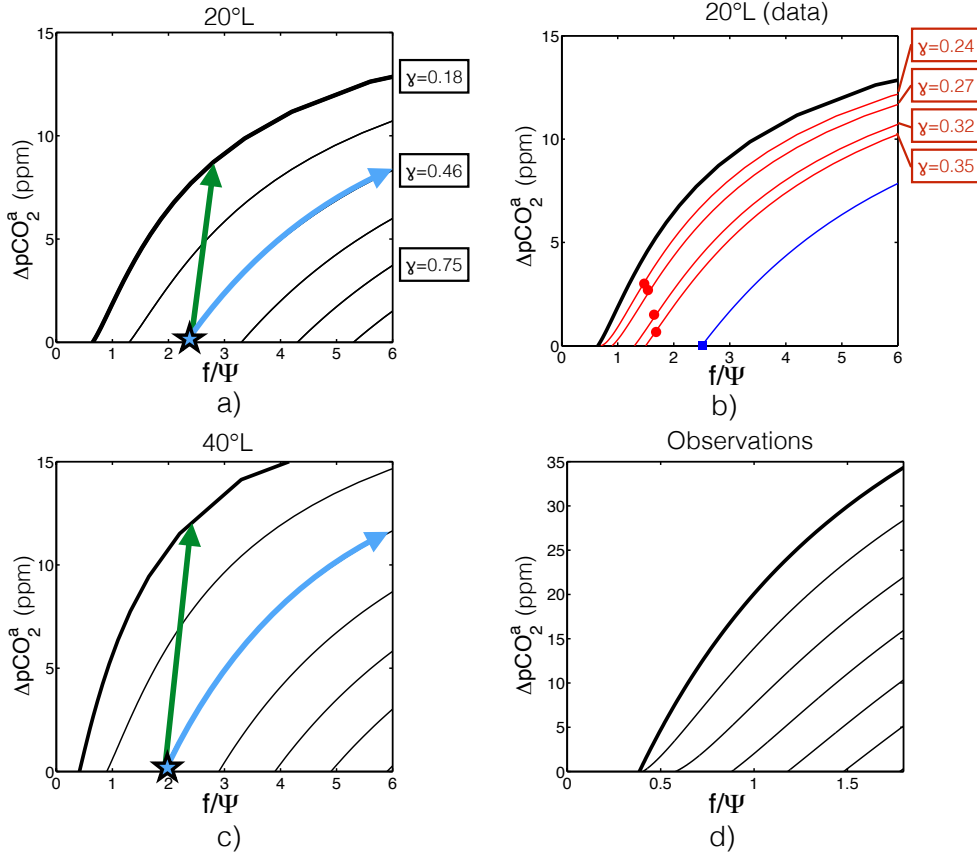


Figure 5.25: The change in atmospheric $p\text{CO}_2$ as a result of an increase in the ratio f/Ψ , with black lines of constant γ and constant f , plotted for a) the 20° longitude model, with panel (b) showing MITgcm data from the 20° longitude Northern salinity restoring perturbation experiments, c) the 40° longitude model, d) N^* as estimated from WOCE by Ito and Follows [2005] and R from Keeling [1973]. In panel (b) the MITgcm control run is shown with a blue square and salinity perturbation experiments on red circles after 5000 years of experiment. The red and blue contours indicate the values of γ associated with the data points. The stars indicate an example control value of f/Ψ , the blue arrow a theoretical line for a new equilibrium given a constant f and γ . The green arrow shows a more realistic line for a new equilibrium with reduced MOC, since f and γ vary.

However, since vertical mixing is linked to the MOC, f is not constant. As the MOC weakens and surface nutrient supply becomes more limited, γ decreases. Allowing for f to vary, the system therefore equilibrates on a different black line, at a different γ , than the one indicated by the blue arrow (see Fig. 5.25a, indicated schematically by the green arrow). The maximum change in atmospheric $p\text{CO}_2$ is then set by the minimum value of γ in the system. This explains why we see maximum changes in $p\text{CO}_2^a$ of less than 10ppm in the model.

We can calculate where the MITgcm data lie on the diagram in Fig. 5.25a by estimating γ in the mixed layer using: $\gamma = N(N + K_N)^{-1}$ [Monod, 1949], where N is the total nutrient concentration $N = N_b + N_p$ and K_N is the half-saturation constant [Monod, 1949] and inferring f/Ψ from the efficiency of the biological pump N^* . The data points (Fig. 5.25b) lie within the region bounded by the grey contours, but the reduction in the overturning causes a strong decrease in γ and therefore an apparent reduction in f/Ψ . One caveat of the approach is the assumption that the biological and preformed responses are decoupled. In reality, they are coupled through C_{dis} , therefore part of the excess upwelled biological carbon mixed in the surface through f will be downwelled again, resulting most likely in a smaller change in $p\text{CO}_2^a$.

Results from a similar model with a domain twice the width, i.e. 40° in longitude, and stronger MOC (6.2 Sv) show a larger change in atmospheric $p\text{CO}_2$ compared to the 20° longitude model. Using eqn. 5.10 with the 40° longitude model reveals that the difference is due to the relative efficiencies of the biological pump (since we saw in section 2.4.1 that surface biology was relatively weak in the 20° in longitude model setup). The 40° longitude model has a higher N^* (at $N^* = 0.17$) and therefore a larger increase in atmospheric $p\text{CO}_2$ can be achieved by reducing the MOC strength (Fig. 5.25c). The difference is therefore not explicitly dependent on the absolute magnitude of the MOC in the control, though a stronger MOC will lead to a more efficient supply of nutrients to the surface, and a higher value of N^* .

Previous studies such as Schmittner et al. [2007] and [Zickfeld et al., 2008] model atmospheric $p\text{CO}_2$ changes up to 30 ppm in response to an MOC shutdown, which is larger than predicted from our simulations. However, Ito and Follows [2005] analyze WOCE⁷ data to estimate N^* to be 0.36, as opposed to 0.16 in our simulations. Using this value gives an upper bound on f/Ψ of 1.8. Using 1.8 for f/Ψ and an estimate of the Buffer Factor R from Keeling [1973] produces the diagram shown in fig. 5.25d, which allows for changes in atmospheric $p\text{CO}_2$ up to 34ppm. While the simple models used to estimate atmospheric $p\text{CO}_2$ changes in response to MOC changes seem to yield reasonable values, the low biological pump efficiency N^* in the present MITgcm model setup causes an underestimation of the $p\text{CO}_2^a$ changes.

⁷World Ocean Circulation Experiment

Diagrams such as Fig. 5.25a-d are useful for estimating the range of atmospheric $p\text{CO}_2$ changes one can expect due to a weakening of the MOC. While the magnitude of the steady state MOC appears in eqn. 5.10 for preformed carbon, the dominant response is given by the regenerated carbon reservoir since the net change in atmospheric $p\text{CO}_2$ is always positive. The atmospheric $p\text{CO}_2$ response is therefore found to be dictated by N^* and therefore $\langle f/\Psi \rangle$. A stronger overturning relative to mixing of biological carbon through f supplies more nutrients to the surface driving a stronger biological pump, which results in a stronger atmospheric response due to MOC changes. All that is required to estimate the range of atmospheric $p\text{CO}_2$ changes are estimates of N^* , nutrient and DIC concentrations and surface ocean $p\text{CO}_2$ in the control state (the ratio f/Ψ can be inferred from eqn. 5.7).

The simple models can also be applied to study the response of atmospheric CO_2 to reductions in the bottom overturning cell. The models shown in Fig. 5.24 separate the MOC, via Ψ , from other vertical processes within the ocean through the mixing term f , so the bottom cell is therefore included within the mixing term f in our formulation. We can study the response of the preformed and regenerated ocean carbon to changes in the bottom cell by calculating the change in atmospheric $p\text{CO}_2$ as f decreases using eqn. 5.10 while holding Ψ constant. Figure 5.26 show the results of the calculation for both the 20° and 40° longitude models. Fig. 5.26a shows that the range of atmospheric $\Delta p\text{CO}_2$ due

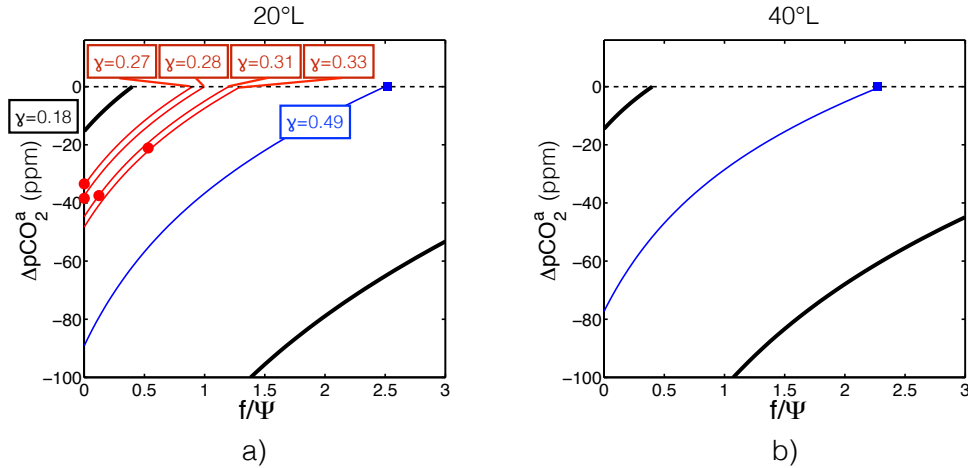


Figure 5.26: The change in atmospheric $p\text{CO}_2$ as a result of a reduction in the ratio f/Ψ , along contours of constant γ and constant f for a) the 20° longitude model and b) the 40° longitude model. Panel (a) includes data from MITgcm control run (blue) and salinity perturbation experiments (red) after 5000 years of experiment. The red and blue contours indicate the values of γ associated with the data points.

to changes in the bottom cell is much larger than the MOC (compare with Fig. 5.25a). Much like the MOC response, the MITgcm data shows that instead of following a curve of fixed γ , γ decreases as f decreases. However, the reduction in γ due to changes in the bottom cell is less than that seen due to MOC changes. In the Northern salinity perturbation experiments where S_N is varied, γ is reduced by about 80% for the MOC shutdown, whereas a shutdown of the bottom cell only causes a reduction of 40% in γ , relative to the control state. The MOC therefore takes a key role in supplying the surface ocean with nutrients. The bottom cell predominantly stores excess biological carbon, which is released into the atmosphere, and excess nutrients which are mostly re-subducted instead of being used up by biology. A reduction in the bottom cell therefore accumulates carbon without affecting surface biology to the same extent as the MOC, resulting in a comparatively larger change in the climate-carbon feedback in $p\text{CO}_2^a$ than with the MOC.

While the importance of bottom cell circulation for the climate has previously been established [Skinner et al., 2010, Toggweiler et al., 2006], the simple box models illustrate the compensation between preformed and biological carbon. The new approach highlights the key differences between the upper and bottom cells in influencing surface biology and atmospheric $p\text{CO}_2^a$, by incorporating the bottom cell into the mixing mechanism f . Moreover, eqn. 5.10 for atmospheric $\Delta p\text{CO}_2$ extends previous studies performed with similar models such as Marinov et al. [2006, 2008b], by allowing explicit calculations of the atmospheric response to changes in the overturning, knowing only a handful of pre-industrial parameters such as N^* and γ without the need to directly estimate f and Ψ .

5.3 Conclusions and climate implications

5.3.1 Implications for anthropogenic carbon

The aim of this chapter was to quantify oceanic carbon inventory changes due to ocean warming in the context of climate change. Firstly, we examined the response of North Atlantic and Southern Ocean $p\text{CO}_2$ to elevated sea surface temperatures. The North Atlantic and Southern Ocean are important for mixing atmospheric carbon anomalies into the deep ocean [Caldeira and Duffy, 2000, Sabine et al., 2004, Khatiwala et al., 2009, Frolicher et al., 2015]. Under warming scenarios, we therefore expect the behaviour of the high latitude ocean $p\text{CO}_2$, $p\text{CO}_2^o$, to affect the anthropogenic carbon uptake.

We find that the North Atlantic $p\text{CO}_2^o$ is relatively constant under increases in surface warming up to 2°C , which agrees with other studies such as Halloran et al. [2015], who suggest that projected long-term trends in the North Atlantic carbon sink are not driven by temperature, but by chemical processes. As atmospheric $p\text{CO}_2$ rises due to emissions of carbon, we therefore expect the North Atlantic carbon sink to be equally efficient at absorbing anthropogenic carbon, C_{ant} , under different degrees of warming. Conversely, the Southern Ocean $p\text{CO}_2^o$ is found to increase as the global ocean warms following an empirical relationship for $p\text{CO}_2$ -temperature solubility [Takahashi et al., 1993]. As the warming increases, the Southern Ocean will most likely become less efficient at absorbing the atmospheric carbon anomaly. To quantify the evolution of the ocean anthropogenic carbon sink as the climate warms, similar anthropogenic carbon experiments as described in Chapter 3 are performed. At the same time as the step forcing in surface temperature restoring is applied, a 260 ppm pulse of carbon is introduced into the atmosphere after which the system is left to equilibrate. The difference in air-sea carbon flux between the anthropogenic pulse experiment and the temperature perturbation-only run then gives the anthropogenic carbon flux $F_{air-sea}^{ant}$. The anthropogenic pulse experiments were only run for 400 years since subsequent differences in anthropogenic carbon uptake between the forcing scenarios are small after this time.

Figure 5.27a shows the time-integrated (cumulative) anthropogenic carbon flux at year 400 in the control run (with no temperature forcing), and Fig. 5.27b shows the change in the cumulative flux in the $1.3\times T$ experiment relative to the milder $1.1\times T$ run (where red indicates ocean uptake and blue indicates ocean outgassing of carbon). The cumulative $F_{air-sea}^{ant}$ in the Southern Ocean is generally reduced due to the stronger warming while the flux in the North Atlantic is stronger. We must take care to consider the atmospheric $p\text{CO}_2$ contribution to the flux, since $p\text{CO}_2^a$ is higher in the $1.3\times T$ experiment due to the background natural carbon response. Therefore, while the North Atlantic $p\text{CO}_2^o$ remains roughly constant, the air-sea carbon flux increases with the strength of the warming. Only a comparison between $1.3\times T$ and $1.1\times T$ experiments is shown, but the same patterns are observed for all simulated amounts of global warming.

The evolution of the anthropogenic carbon flux can be quantified by considering the cumulative uptake. Figure 5.27c shows the cumulative anthropogenic carbon air-sea flux integrated spatially over both the Southern Ocean and North Atlantic after 400 years of

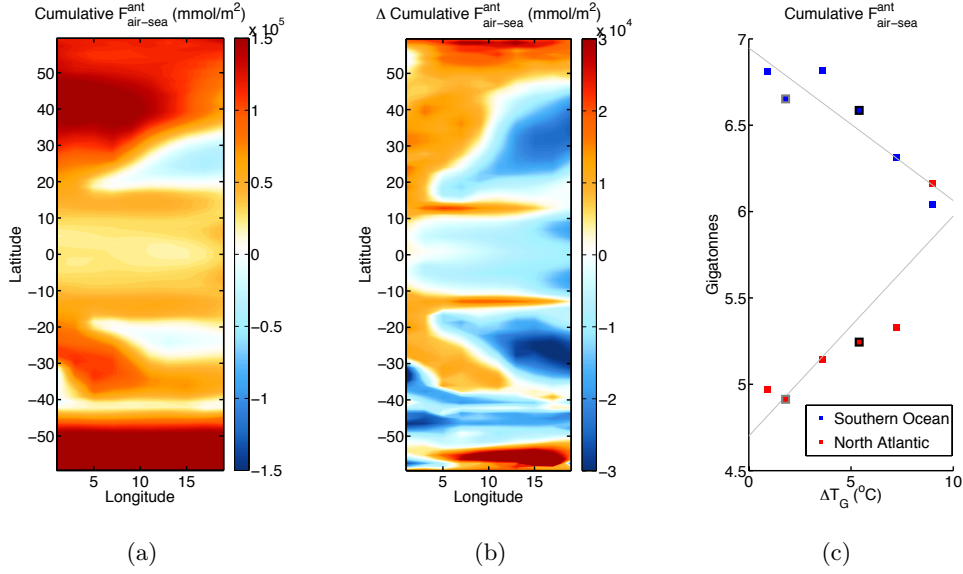


Figure 5.27: a) Cumulative anthropogenic carbon flux $F_{air-sea}^{ant}$ in MITgcm at year 400 after a pulse emission without any temperature changes, b) difference in the cumulative $F_{air-sea}^{ant}$ after 400 years between $1.3 \times T$ and $1.1 \times T$ experiments; c) Cumulative $F_{air-sea}^{ant}$ over the North Atlantic (red) and Southern Ocean (blue) after 400 years as a function of global temperature anomaly ΔT_G . The data points with the grey outline correspond to the $1.1 \times T$ experiment and those with the black outline, to the $1.3 \times T$ experiment.

experiment as a function of the global mean temperature anomaly ΔT_G . As the global mean temperature forcing increases, the cumulative flux in the Southern Ocean reduces whereas the North Atlantic cumulative flux increases at a similar rate (except for the extreme value of $\Delta T_G = 9^\circ\text{C}$ global warming). Figure 5.27c also shows that as the warming increases, the difference in anthropogenic carbon uptake between the North Atlantic and Southern Ocean is reduced.

The relative responses of the Southern Ocean and North Atlantic C_{ant} sinks are roughly equal and opposite. The implication is that the rate of ocean uptake of anthropogenic carbon is roughly constant under different amounts of warming. Figure 5.28a shows the time evolution of the total oceanic C_{ant} , converted to equivalent atmospheric $p\text{CO}_2$ taken up, for a range of global mean warming up to $\Delta T_G \approx 10^\circ\text{C}$ (equivalent to the $1.5 \times T$ experiment). The spread between the total uptake at year 400 is 12 ppm, which represents a 5% difference from the lowest warming case. The largest spread occurs in the first 150 years due to the different timescales of North Atlantic carbon adjustment (Fig. 5.28a).

The consistency in C_{ant} uptake across the range of global mean warming in Fig. 5.28a

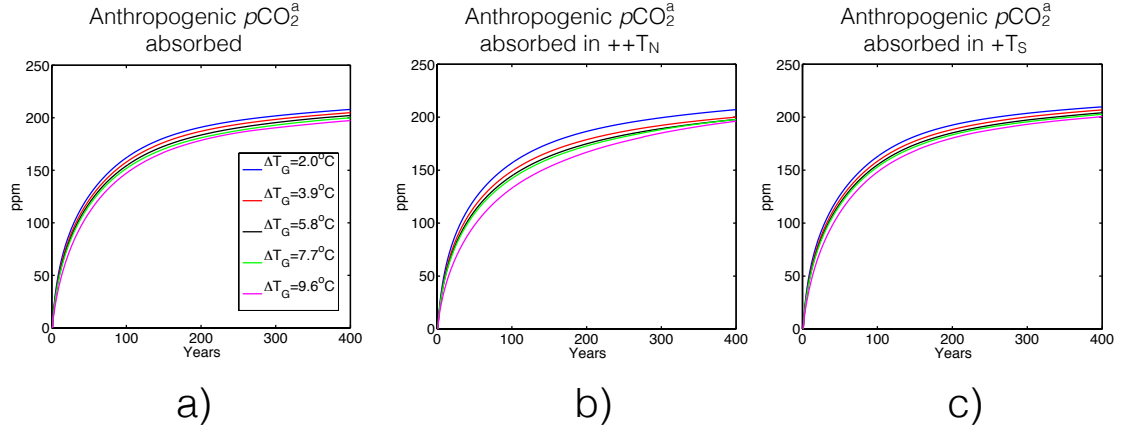


Figure 5.28: Atmospheric $p\text{CO}_2^{\text{ant}}$ taken up by the ocean for different amounts of global mean warming ΔT_G , in ppm from a total of 260ppm, for the a) base warming, b) ‘++ T_N ’ and d) ‘+ T_S ’ scenarios.

is also seen in similar anthropogenic pulse runs for the ‘+ T_N ’, ‘++ T_N ’ (Fig. 5.28b) and ‘+ T_S ’ (Fig. 5.28c) experiments. Northern warming in the ‘+ T_N ’ and ‘++ T_N ’ scenarios results in a decrease in the upper cell of the overturning, as well as a decrease in the meridional SST gradient. Both reduce the cooling heat flux experienced by waters advected by the sub-tropical western boundary current which studies identify as an important factor in determining the North Atlantic C_{ant} uptake⁸ [Alvarez et al., 2003, Macdonald et al., 2003, Roson et al., 2003, Quay et al., 2007, Perez et al., 2013]. A weakening of the MOC decreases C_{ant} downstream of the sub-tropical gyre western boundary current, as observed in the model (fig. 5.29b) and so North Atlantic C_{ant} is reduced, while Southern Ocean C_{ant} is increased. Consequently, the North Atlantic and Southern Ocean anthropogenic carbon uptake compensate such that the overall uptake of C_{ant} remains roughly constant as the system warms.

In sections 5.1 and 5.2 we saw that the low latitude warming determines 75% of the natural atmospheric $p\text{CO}_2$ response over centennial timescales. The low latitudes also show the strongest air-sea flux of anthropogenic carbon within the first decade of the pulse. However, due to the shallow circulation, no significant amount of anthropogenic carbon is permanently stored there. After the first decade, the anthropogenic carbon is either advected polewards or released into the atmosphere on the eastern side of the

⁸More than 50% of the anthropogenic carbon in the North Atlantic sub-polar gyre was transported into the region via the upper cell of the MOC

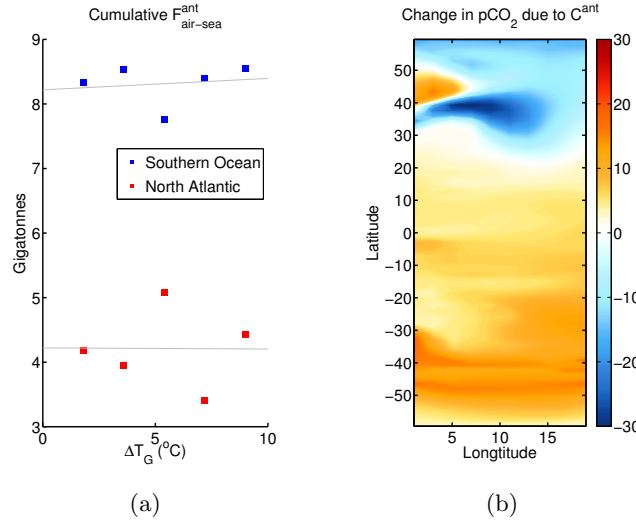


Figure 5.29: a) Cumulative $F_{air-sea}^{ant}$ flux over the North Atlantic (red) and Southern Ocean (blue) at year 400 as a function of global temperature anomaly $\Delta \overline{T}_G$ in the $+T_N$ experiment, b) The difference in the anthropogenic $p\text{CO}_2^a$ signal between the $+T_N$ and base warming experiments, in ppm.

gyres (see Fig. 5.27a). As the oceans warm and the meridional temperature gradient increases, the heat flux out of the ocean along the western boundary increases causing an increased uptake of carbon. However, the return flow in the eastern part of the gyre will experience an increased heat flux *into* the ocean causing strong outgassing of the anthropogenic carbon. The total zonal sum of the uptake is therefore small in the mid-latitudes compared to the high latitudes.

In response to warming, the full atmospheric carbon response (natural + anthropogenic) will therefore be dictated by the low latitudes release of natural carbon, since the high latitude response to temperature buoyancy forcing is compensated between the Northern and Southern high latitudes.

5.3.2 Implications for natural carbon

The low-latitude natural atmospheric $p\text{CO}_2$ feedback is determined by the global mean temperature on centennial timescales. The solubility trend in $p\text{CO}_2^a$ accounts for around 75% of the signal, while the reduction in surface biology accounts for the remaining 25%. The effects of dynamical adjustment were found to be negligible over centennial timescales; this was tested by fixing ocean dynamics and perturbing boundary conditions. As a result of changing dynamics, the change in saturated carbon C_{sat} due to a redistribution of

heat is compensated by an almost equal and opposite change in the disequilibrium carbon component, C_{dis} , due to a slowing of mode water formation.

An important region in the centennial carbon compensation is the Southern Ocean, where the Antarctic Intermediate Water (AAIW) is formed. The disequilibrium carbon reservoir, C_{dis} , is generally large in AAIW [Lauderdale et al., 2013] so any change in AAIW formation will affect global C_{dis} concentrations. An important factor in the formation of AAIW is the Southern Ocean residual circulation, which weakens only slightly due to the warming. In the presence of stronger eddies, the bolus circulation will respond more strongly to changes in horizontal density gradients. We can mimic a stronger eddy response by increasing the value of the eddy diffusivity, κ_{GM} , manually in the MITgcm model. By increasing κ_{GM} , buoyancy perturbations will cause a stronger response in the bolus circulation to surface warming and therefore the residual circulation will be altered. A perturbed residual circulation affects how the C_{dis} reservoir responds to surface warming, breaking the dynamical balance between C_{dis} and C_{sat} .

Figure 5.30 shows the atmospheric $p\text{CO}_2$ anomaly after 400 years for the same warming experiments as shown in Fig. 5.14a, along with a base warming scenario where we double the strength of the eddy response by doubling the eddy diffusivity κ_{GM} . The increase in eddy forcing leads to a reduction in the residual circulation and a reduced change in $p\text{CO}_2$ compared to all other warming experiments. The results show that strong changes in local dynamics induced by additional climate forcing such as wind stress could break the balance of dynamical carbon compensation.

These new results provide insights into the global feedback since understanding ocean anthropogenic carbon alone will not provide a full picture of the carbon feedback. Studies such as Plattner et al. [2001] find that the ocean carbon feedback to global warming is dominated by the low latitudes on timescales of longer than a century and the study presented here suggests that this is due to the natural carbon feedback rather than any anthropogenic carbon signal. Moreover, it is shown that the response is independent of dynamics over centennial timescales. Changes in biological productivity due to anthropogenic carbon were not addressed, such as ocean acidification at higher concentration of dissolved CO_2 , since such feedbacks are not included in the MITgcm. Global mean biological productivity will likely decrease as the global temperature rises [Caldeira and

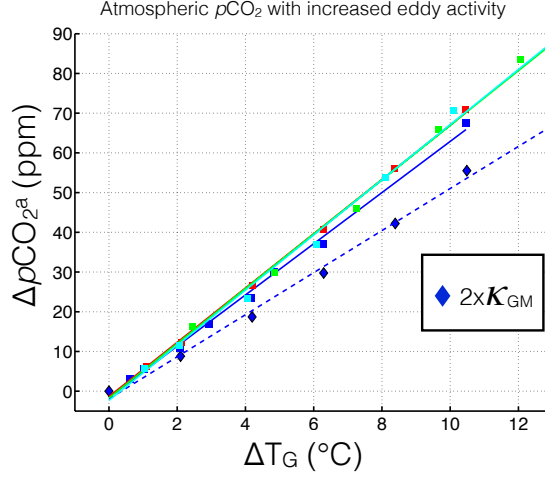


Figure 5.30: Change in atmospheric $p\text{CO}_2$ in MITgcm as a function of global mean sea surface temperature anomaly ΔT_G after 400 years of experiment. Squares indicate model data from warming experiments according to the colour-coded profiles in Fig. 5.9 and the solid lines are the corresponding linear trends. The blue diamonds indicate warming experiments with base warming scenario and enhanced eddy activity by doubling the eddy diffusivity parameter κ_{GM} .

Wickett, 2003, Orr et al., 2005, Solomon et al., 2007, Doney et al., 2009], though the climate impact of such changes is uncertain. The simplicity of the biological model is therefore a limitation in the presented study.

Compensations between ocean carbon reservoirs limit the overall atmospheric $p\text{CO}_2$ changes over longer timescales as well. Over millennial timescales, an increase in stratification due to global surface warming causes an accumulation of regenerated carbon C_{reg} at depth which reduces the overall atmospheric $p\text{CO}_2$ response to warming. Similarly, perturbations in the large scale overturning result in a partial balance between C_{dis} and regenerated carbon. Using idealised salinity perturbation experiments and box models, a decrease in the upper overturning cell results in an increase in $p\text{CO}_2^a$ due a loss in C_{reg} . However, the response in $p\text{CO}_2^a$ to a decreasing overturning is generally small (5 ppm in the MITgcm, 30ppm in the real ocean) due to the partial compensation of C_{dis} with C_{reg} . For example, in the $S_N = 33$ psu experiment, corresponding to an MOC of 2.1 Sv, if there was no compensation between C_{reg} by C_{dis} , the atmospheric $p\text{CO}_2$ change would be 15 times larger since the regenerated carbon reservoir is decreased by 10%. In this case, atmospheric $p\text{CO}_2$ change could be of order 80 ppm, instead of 5 ppm, in response to MOC changes in the MITgcm, or up to hundreds of ppm in the real ocean.

A similar balance occurs as a result of a decrease in the bottom overturning cell. However, the increase in C_{dis} is larger than the reduction in C_{reg} because the bottom cell mainly ventilates old regenerated carbon rather than supplying the surface with necessary nutrients for biological production. Studies such as Marinov et al. [2006] allude to such a split by recognising that the sub-polar and sub-Antarctic regions of the Southern Ocean play different roles in the ocean carbon cycle, though they do not elaborate on the role of both overturning cells.

Using the insights from the salinity perturbation experiments and the box models, the millennial atmospheric $p\text{CO}_2$ changes in Fig. 5.14 can be interpreted. Previously, it was shown that the boundary conditions make little impact on atmospheric $p\text{CO}_2$ on centennial timescales. On millennial timescales, a reduction in the sensitivity of the $p\text{CO}_2^a$ anomaly of 10-15 ppm in the Southern warming experiments is modelled, relative to the base warming case (cyan line in Fig. 5.14b). Such a change is consistent with the Southern salinity perturbation experiments and is mainly driven by the accumulation of regenerated carbon in the bottom overturning cell.

The North Atlantic warming experiments are expected to show an elevated sensitivity of $p\text{CO}_2^a$ to ΔT_G of up to 5ppm in the most extreme cases of polar warming. However, Fig. 5.14b shows a decrease in the sensitivity of $p\text{CO}_2^a$ of up to 35ppm. Section 5.2.2 shows that the $p\text{CO}_2^a$ contribution due to dynamical changes is negative and becomes significant after a few centuries. The differences in the saturated and disequilibrium carbon reservoirs between the $++T_N$ Northern warming and the base warming scenario are small and mostly compensate each other when averaged over the whole basin. However, the difference in the regenerated carbon reservoir is comparatively much larger (Fig. 5.31). The change in $p\text{CO}_2^a$ due to the dynamics is therefore largely associated with the isolation of biological carbon in the Northern Hemisphere. To some extent this effect is seen in the base warming scenario, which results in the reduction in the $p\text{CO}_2^a$ anomaly between 400 and 3000 years in Fig. 5.14, though it is not as pronounced.

In the Northern salinity perturbation experiments, there is a similar isolation of C_{reg} in the Northern hemisphere. However, in the salinity experiments, the increase in C_{reg} is over-compensated by the reduction in surface biology in the Southern hemisphere and consequently, the reduction of C_{reg} in the bottom overturning cell (see Fig. 5.19d). In

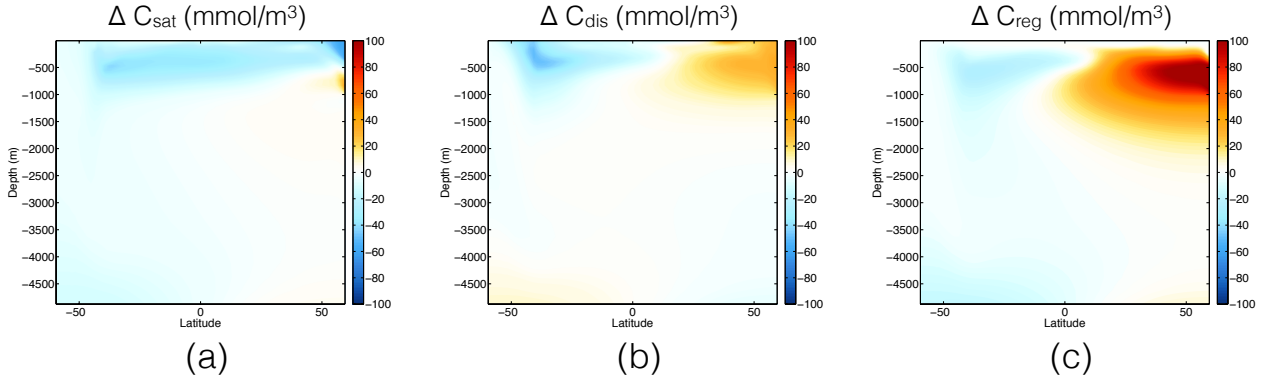


Figure 5.31: Zonal mean difference in a) saturated, b) disequilibrium and c) regenerated carbon after 3000 years between the $++T_N$ and base warming MITgcm warming scenarios in the $1.4\times T$ experiment.

all warming experiments however, C_{reg} accumulates in the slowing bottom overturning cell, which is weakened because of an increase in Southern channel stratification (as in Fig. 5.22). Due to the presence of mid-latitude warming in all warming scenarios, the increase in stratification in $++T_N$ experiments affects surface biology less than the salinity perturbation experiments. For example: consider a weakening of the MOC of 2 Sv. The 2 Sv reduction can be achieved by either a Northern boundary salinity perturbation or by increasing the Northern boundary temperature T_N between two warming experiments. In the case of salinity forcing, the reduction in Southern hemisphere biology is 35%. However, Southern hemisphere biology is only reduced by 14% in the case of increased Northern warming. Therefore, Northern warming gives a negative atmospheric pCO_2 anomaly, unlike an isolated salinity perturbation. As the Northern boundary warming increases, the reduction in pCO_2 also increases. In the case of strongest Northern boundary warming, $++T_N$, the dynamical changes reduce the net pCO_2 change by 60% and compensate for a large proportion of the changes in C_{sat} .

On millennial timescales, the negative dynamical feedback of warming on pCO_2 is therefore strengthened by polar warming and the C_{reg} reservoir compensates the changes in saturated carbon, C_{sat} . In case of strong polar warming, the compensation can lead to small perturbations in atmospheric pCO_2 of less than 10 ppm for $8^\circ C$ of global mean surface warming, which has not been addressed in similar studies such as Joos et al. [1999]. The high-latitude warming-induced compensation is relevant for the real ocean since both observations and coupled climate model point towards a polar amplification of sea surface

temperatures in the contemporary and future ocean [Manabe and Stouffer, 1980, Holland and Bitz, 2003, Reynolds et al., 2007, Timmermans and Proshutinsky, 2014].

The dominance of the isolation of C_{reg} in the Northern hemisphere compared to the Southern hemisphere C_{reg} anomaly in the warming experiments demonstrates the non-linearity of the system. While isolated perturbation experiments, such as the salinity experiments, can give valuable insight into the physical mechanisms setting the carbon cycle response due to particular forcings, the combined result is often not a simple linear superposition (e.g. Schwinger et al. [2014]). The strength of the non-linear response in $p\text{CO}_2^a$ can be larger than predicted from the individual perturbation experiments and theoretical predictions.

A caveat of the present study is the simplicity of the model, particularly the domain bathymetry. While the MITgcm model simulates the important features needed for this study, namely a large scale overturning circulation, the strength of the MOC is relatively weak and the bottom overturning cell is relatively large compared to the real ocean. However, the goal of the study is to give estimates of the magnitude of the warming-induced climate-carbon feedback and the mechanisms causing the feedback. To that end, idealised geometry and forcing scenarios are employed. For detailed estimates of the future climate-carbon feedback and associated errors, a more realistic geometry, biological model and forcing scenarios are required.

The overall message is that ocean warming results in both thermodynamic and dynamic changes which result in large perturbations in the individual ocean carbon reservoirs. If uncompensated, atmospheric $p\text{CO}_2$ anomalies could be of the order of the anthropogenic CO_2 perturbation. However, dynamical perturbations result in compensation between the different carbon reservoirs of saturated, disequilibrium and regenerated carbon. On centennial timescales, dynamics has a negligible influence on atmospheric CO_2 since dynamical changes in C_{sat} are balanced by changes in C_{dis} . Over millennia, strong polar warming can cause non-linear effects, isolating C_{reg} such that it compensates the loss of C_{sat} due to warming. Ultimately, the dynamical carbon compensation across the ocean basin results in a certain level of stability of the ocean carbon system in response to climate-change induced warming. The natural ocean carbon feedback due to surface warming on atmospheric $p\text{CO}_2$ is therefore limited, despite the amount of carbon stored within the ocean.

In order for larger atmospheric $p\text{CO}_2$ changes to be achieved, an isolated change in one of the carbon reservoirs is required. Such a change is unlikely to be achieved through adjustments in the large-scale dynamics. It could be achieved by altering the efficiency of surface biology γ which would only affect regenerated carbon. Additionally, local high latitude processes such as the enhanced eddy activity in the Southern Ocean could alter the extent of the carbon compensation.

Summary of carbon predictions in past, present and future climates

6.1 Anthropogenic ocean carbon and heat storage

In previous chapters, we quantified the climate-carbon feedback induced by temperature and wind stress perturbations on atmospheric $p\text{CO}_2$. In these perturbation scenarios, dynamical changes induced by temperature and wind forcing alter not only the carbon storage but the ocean heat content as well. While ocean heat content changes are important for ocean carbon storage, the rates and patterns of carbon and heat uptake can be vastly different resulting in different timescales of radiative forcing and delays in global warming (see section 2.1). Simple models often assume similar mechanisms and timescales for global mean air-sea heat and carbon fluxes. However, observations and coupled model simulations show disparities between the ocean uptake of carbon and heat [Sabine et al., 2004, Levitus et al., 2005, Khatiwala et al., 2009, Winton et al., 2013, Goodwin et al., 2015]. For example, the Atlantic ocean heat and carbon anomalies in the HadGEM2-ES $2\times\text{CO}_2$ scenario (section 2.2.2) show clear differences and chapter 5 showed that carbon solubility only partly accounts for the atmospheric CO_2 change over multiple timescales. In this section, we extend our analysis by examining the impact of circulation on the meridional transport of heat and carbon. By drawing direct comparisons between heat and carbon transport, we aim to gain further understanding and context of our results from chapters 2 to 5. Specifically, we will attempt to estimate climate change-induced shifts in ocean circulation and heat storage using anthropogenic carbon storage.

6.1.1 Impact of circulation changes on heat and carbon transport

The meridional transport of heat and carbon is a useful indicator of the role of dynamics in redistributing the temperature and dissolved inorganic carbon within the ocean. We can decompose each variable: dissolved inorganic carbon, DIC , temperature, T , and meridional Eulerian velocity, v , into a mean state, $\overline{(\dots)}$, and a perturbation, $(\dots)'$, induced under a forced experiment, such that $DIC = \overline{DIC} + DIC'$, $T = \overline{T} + T'$ and $v = \overline{v} + v'$, respectively. The meridional transport of DIC, $\mathcal{Q}_C$, and heat, $\mathcal{Q}_H$, integrated between

longitudes l_1 and l_2 are given by:

$$\mathcal{Q}_C = \int_{l_1}^{l_2} v \, DIC \, dx = \int_{l_1}^{l_2} (\bar{v} + v') (\overline{DIC} + DIC') \, dx, \quad (6.1)$$

$$\mathcal{Q}_H = \int_{l_1}^{l_2} \rho_0 C_p v \, T \, dx = \int_{l_1}^{l_2} \rho_0 C_p (\bar{v} + v') (\bar{T} + T') \, dx, \quad (6.2)$$

where ρ_0 is the reference density and C_p is the specific heat capacity. For this study, we consider the zonally integrated meridional transport over the Atlantic Ocean and Southern Ocean between -70° and 30°E in HadGEM2-ES. Using the HadGEM2-ES $2\times\text{CO}_2$ experiment, we can compute how the carbon $\mathcal{Q}_C$ and heat $\mathcal{Q}_H$ transports are altered due to both increased DIC and temperature and a change in the Eulerian velocity v .¹ The mean state is defined as the control mean and the perturbation is the difference between the average of the last 10 years of the $2\times\text{CO}_2$ experiment and the control mean.

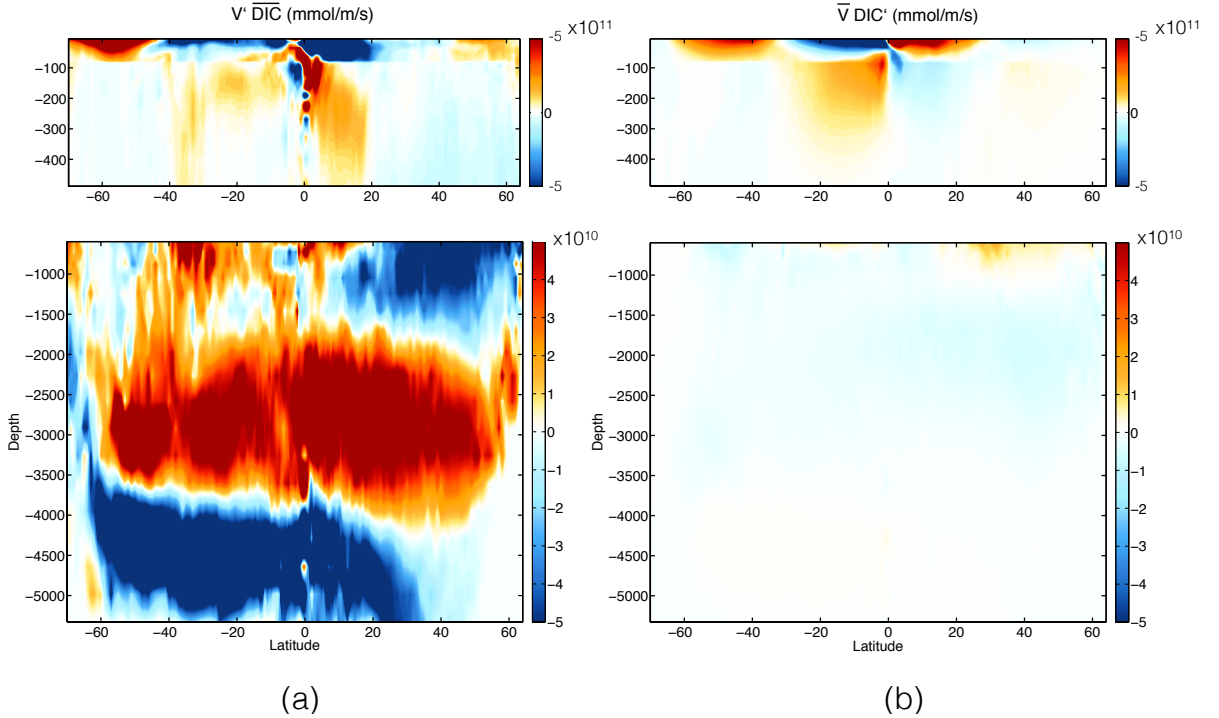


Figure 6.1: Zonally integrated meridional DIC transport anomalies as a function of depth and latitude: a) $v' \overline{DIC}$ and b) $\bar{v} DIC'$ in the $2\times\text{CO}_2$ experiment. The top row shows the upper 500m and the bottom row shows the transport below 500m.

Considering the meridional transport of DIC , $\bar{v} \overline{DIC}$ is the control mean DIC transport so we will focus on the zonally integrated first order cross terms $v' \overline{DIC}$ and $\bar{v} DIC'$ shown in Fig. 6.1. Both $v' \overline{DIC}$ and $\bar{v} DIC'$ contribute to an increase in northward surface

¹Note that we do not have bolus velocities available as an output for HadGEM2-ES so this study will only consider Eulerian transport.

transport in the Southern Ocean, while in the North Atlantic sub-polar gyre (40° - 60° N) the two contributions oppose each other. However, in each region the magnitude of the mean DIC transport due to the anomalous velocity anomaly $v'\overline{DIC}$ is twice as large as the DIC anomaly advected by the mean $\bar{v}DIC'$. Therefore, the anomalies in the surface Ekman velocity, particularly in the Southern Ocean, are mostly responsible for the increased northwards DIC transport [Ito et al., 2010]. Below 500m, the values of $v'\overline{DIC}$ and $\bar{v}DIC'$ are an order of magnitude lower than that at the surface. Nonetheless, $v'\overline{DIC}$ is still relatively more dominant at depth. Throughout the entire basin therefore, the changes in DIC transport are predominantly due to the anomalous circulation $v'\overline{DIC}$.

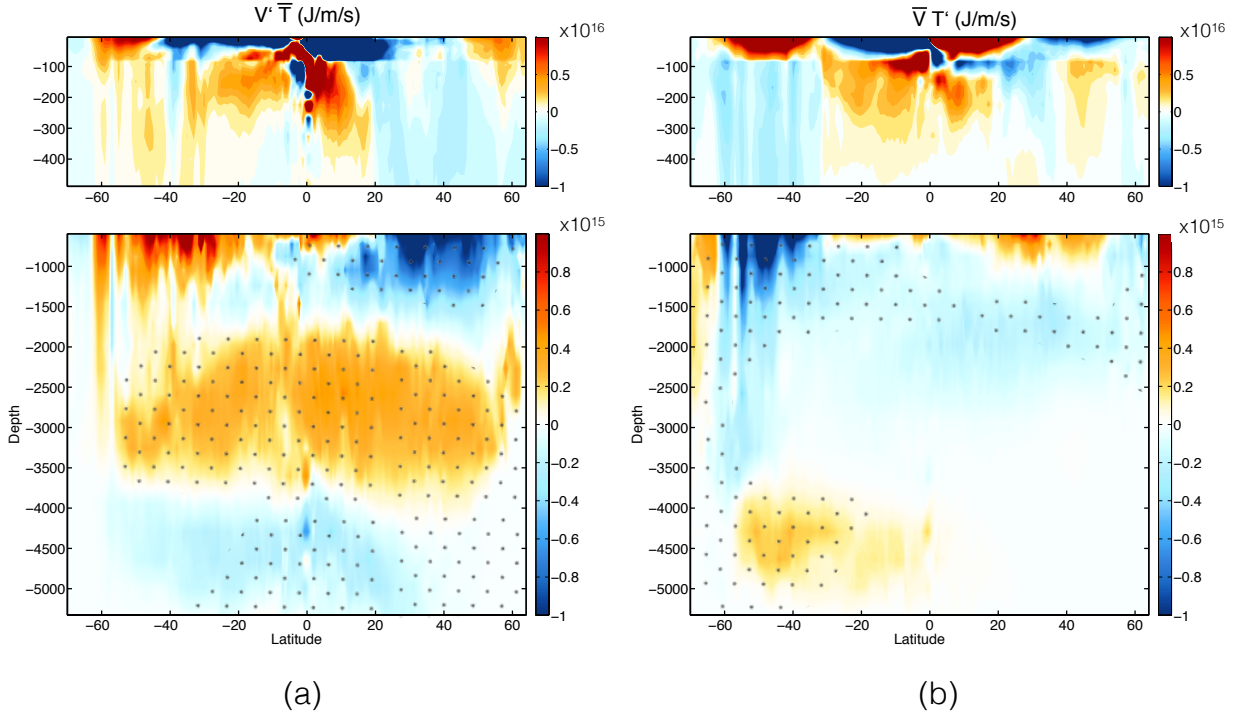


Figure 6.2: Zonally integrated meridional heat transport anomalies: a) $v'\overline{T}$ and b) $\bar{v}T'$ in the $2\times\text{CO}_2$ experiment. The top row shows the upper 500m and the bottom row shows below 500m. The stippling in bottom sections indicate which contribution dominates.

In contrast, the decomposition for the heat transport $\mathcal{Q}_H$ is shown in Fig. 6.2. At the surface, the heat anomaly advected by the mean circulation, $\bar{v}T'$, is larger than $v'\overline{T}$ but only by a factor of 2. In the North Atlantic, both $\bar{v}T'$ and $T'\bar{v}$ are comparable in magnitude. The two contributions to the anomalous meridional heat transport are once again opposite signs because there is a reduction in the background southward velocity field within the sub-polar gyre. In the Tropics, both $\bar{v}T'$ and $T'\bar{v}$ are of similar magnitude. At depth, the magnitudes of the transport anomalies are lower than the surface, however, we note

that $\overline{T}v'$ and $T'\overline{v}$ are dominant in different regions (indicated by the stippling in Fig. 6.2). Regions in which $\overline{T}v'$ dominates at depth are mostly a consequence of the shallowing of the MOC and expansion of the bottom overturning cell. By and large however, the two components of $T'\overline{v}$ and $\overline{T}v'$ are comparable in magnitude so that both are important for the overall anomalous heat transport throughout the ocean.

These results apply not only to $2\times\text{CO}_2$ scenario, but also to more extreme cases such as the $4\times\text{CO}_2$ experiment. Fig. 6.3 shows the response of the Southern Ocean and North Atlantic surface transport of heat and carbon due to the two different CO_2 forcings, including the projected change at $4\times\text{CO}_2$ based on a linear extrapolation from the $2\times\text{CO}_2$ response.

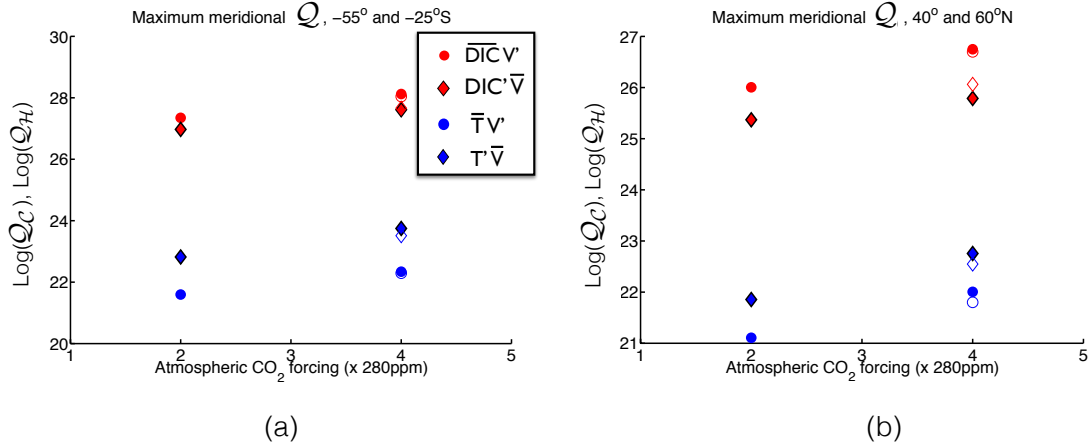


Figure 6.3: Maximum meridional ocean heat and carbon transport in the upper 500m, a) for the Southern Ocean, 55° and 25°S , and b) the North Atlantic sub-polar gyre, 40° and 60°N as a function of atmospheric CO_2 forcing. The filled circles indicate the contribution from the anomalous velocity while the diamonds indicate the contribution from DIC or T anomaly. Red corresponds to the DIC transport and blue to the temperature. The empty circles and diamond show the projected linear increase at $4\times\text{CO}_2$ from a doubling of the $2\times\text{CO}_2$ values. The units on the y-axis are $\log(\text{mmol kg}^{-1} \text{ m}^2 \text{ s}^{-1})$ for DIC and $\log(\text{J m}^2 \text{ s}^{-1})$ for temperature.

The heat and carbon transports at $4\times\text{CO}_2$ deviate from the extrapolated linear response, as the forcing is strong enough to induce non-linear effects which amplify the response of the transport. The deviation from the linear estimates in the Southern Ocean are relatively small compared to the North Atlantic subpolar gyre. Only the contribution of $\text{DIC}'\overline{v}$ is overestimated by the linear extrapolation, which is partly due to surface warming-reduced solubility and chemical changes (particularly in the North Atlantic). For

the heat transport, both contributions, $T'\bar{v}$ and $\bar{T}v'$, show a strong non-linear amplification. Overall, the gap between $\overline{DIC}v'$ and $DIC'\bar{v}$ widens with the strength of the forcing due to the non-linearities. The gap between $T'\bar{v}$ and $\bar{T}v'$ does widen in the Southern Ocean, but remains roughly constant in the North Atlantic. The dominance of surface circulation anomalies for carbon transport therefore increases with the forcing, though the difference in $T'\bar{v}$ and $\bar{T}v'$ does not widen convincingly everywhere.

The global importance of the $\overline{DIC}v'$ is a result of the large scale circulation changes having little effect on the concentration of DIC. By contrast, the magnitude of temperature anomalies results in a comparable magnitude of $T'\bar{v}$ and $\bar{T}v'$. The relevance for estimating climate feedbacks is profound: only because $DIC'\bar{v}$ is small, especially at depth, can we derive the scaling in chapter 3. The scaling estimates changes in $\overline{DIC}v'$ (or rather, the 3D transport anomalies $\overline{DIC}\mathbf{u}'$) and assumes $DIC'\bar{v}$ in the deep ocean does not change.

To validate our previous assumptions and to compare carbon and heat transports, Fig. 6.4 shows the DIC and transport anomalies in the $1.5\times\tau_{SO}$ experiment after 400 years with an anthropogenic pulse emission of 260 ppm (but no increase in radiative forcing), as well as the control state background fields. The magnitude of the $DIC'\bar{v}$ anomaly (panel b in Fig. 6.4), even with the anthropogenic pulse, is negligible in comparison to $\overline{DIC}v'$ (panel d in Fig. 6.4) and the background $\bar{v}\overline{DIC}$ field (panel f in Fig. 6.4). However, both temperature transport anomalies $T'\bar{v}$ and $\bar{T}v'$ (panels a and c in Fig. 6.4, respectively) are within the same order of magnitude. Adjustments in ocean circulation therefore have a strong impact on temperature anomalies so constructing a mixed layer scaling for temperature would not be successful, illustrating the need for different approaches in determining ocean and heat carbon feedbacks, as discussed in section 2.1. The net change in ocean heat content in the 400-year $1.5\times\tau_{SO}$ experiment is $3.1\times 10^{23}\text{J}$, equivalent to the change induced in the base surface warming experiment of $\Delta T_G = 4.5^\circ\text{C}$. Similar changes are observed for any of the wind stress perturbation experiments.

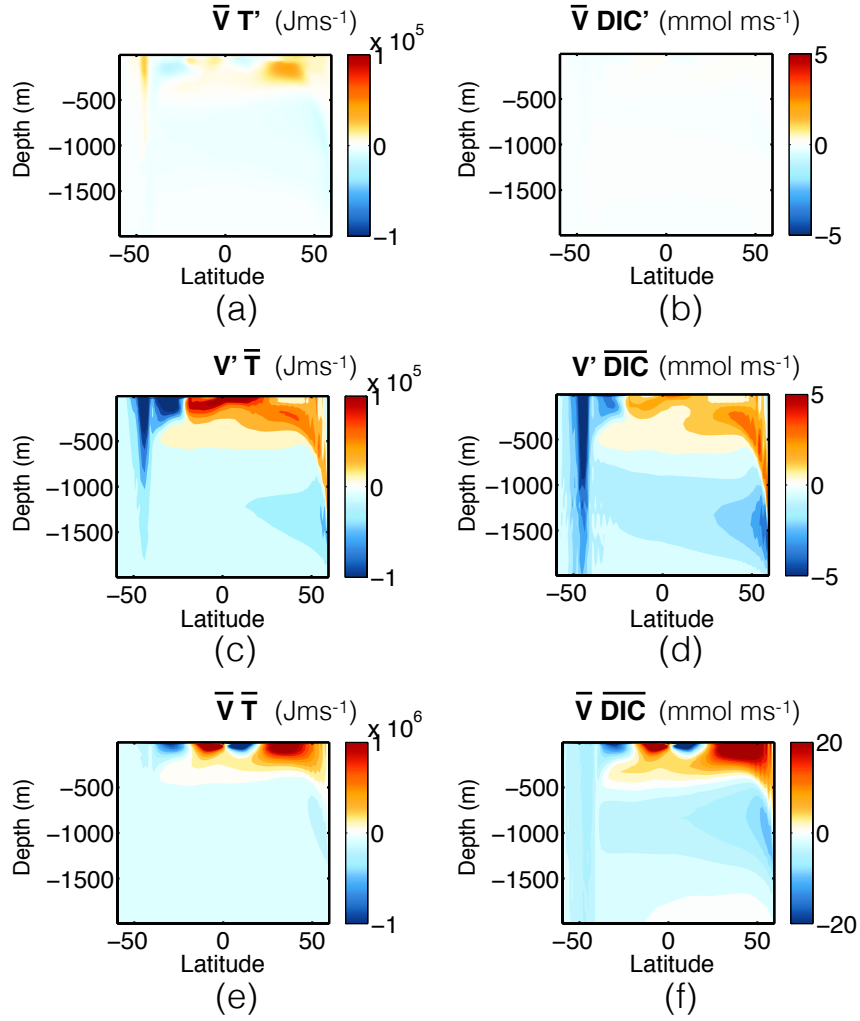


Figure 6.4: Zonal mean meridional transport anomalies after 400 years in the $1.5\tau_{SO}$ experiment with anthropogenic pulse emission of 260 ppm as a function of depth and latitude: a) $\bar{v} T'$, b) $\bar{v} DIC'$, c) $v' \bar{T}$, d) $v' \overline{DIC}$. Panels e-f show the background control mean transport: e) $\bar{v} \bar{T}$ and f) $\bar{v} \overline{DIC}$.

The experiments show that dynamical changes have significant impacts on ocean heat storage. The temperature anomalies themselves change buoyancy gradients and feedback onto the transport, air-sea flux and overall heat uptake. To assess the magnitude of the circulation-induced heat uptake anomaly, we can employ the fixed dynamics MITgcm experiments from chapter 5 to calculate the change in ocean heat uptake that results from the buoyancy-induced circulation changes. Figure 6.5 shows the ocean heat anomaly after 400 years in the base surface warming experiments for both the fixed dynamics (magenta) and freely evolving (blue) MITgcm experiments. Figure 6.5a shows that the dynamical adjustments reduce global heat uptake by 28% (while the data after 400 years is shown,

the relative impact of the dynamics remain constant with time), which is consistent with the similar estimate of 27% by Winton et al. [2013] using a more realistic model setup. To accurately calculate changes in ocean uptake, we therefore need to know how the circulation field adapts to anthropogenic forcings.

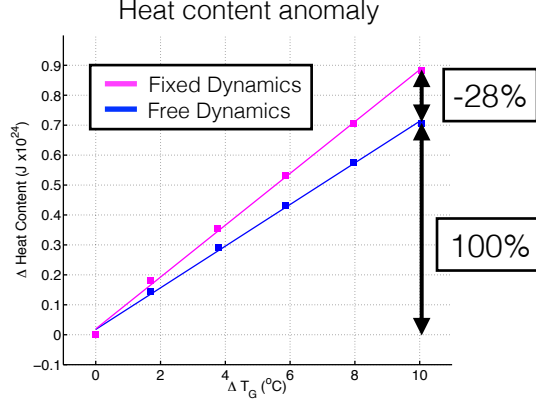


Figure 6.5: Ocean heat content anomaly as a function of global mean sea surface temperature after 400 years in MITgcm with free (blue) and fixed (magenta) dynamics. Squares show MITgcm data and the solid lines are linear fits.

6.1.2 Reconstructing ocean heat anomalies

The MITgcm fixed circulation experiments show that the circulation-induced temperature anomalies account for 28% of the global change in heat uptake. While knowing the magnitude is useful, there is also much to be gained from knowing the spatial distribution of the circulation-induced temperature anomalies. The circulation-induced anomalies re-distribute heat within the ocean, potentially cooling and compensating surface-forced warming [Winton et al., 2013].

For a forced CO_2 experiment, as discussed before, the terms $\bar{T}$ and $\bar{\mathbf{u}}$ correspond to the control (unperturbed) mean state while T' and $\mathbf{u}'$ are the perturbed temperature and full 3D velocity field, respectively. If we consider a perturbed experiment in which the circulation is fixed to its preindustrial value, then $\mathbf{u}' = 0$. In the fixed circulation experiment, the temperature perturbations are only redistributed by the state steady velocity leading to different patterns of temperature anomalies compared to a freely evolving perturbed experiment. The full (freely evolving) temperature anomaly field T' is therefore $T' = T'_{free} = T'_{fix} + T'_{dynamics}$, where T'_{fix} is the passive component advected by the

steady state velocity $\bar{\mathbf{u}}$ induced by a change in radiative forcing at the surface only (similar to the passive anomaly temperature in Banks and Gregory [2006]) and $T'_{dynamics}$ is the dynamically-induced temperature perturbation. The spatial patterns of T'_{fix} are similar to passive tracers such as anthropogenic carbon C_{ant} . To illustrate the difference and similarities between T'_{fix} , T'_{free} and C_{ant} , consider the $1.1 \times T$ MITgcm base warming experiment with an anthropogenic carbon pulse (the $1.1 \times T$ experiment was chosen because it has a 2°C global mean warming anomaly, similar to the $2 \times \text{CO}_2$ HadGEM2-ES experiment).

To compare the patterns of temperature and carbon anomalies, the anomalies have been normalised with respect to the domain-averaged value of the anomaly at time t . For example, the normalised anomaly, T'^n_{fix} , is defined as:

$$T'^n_{fix}(x, y, z, t) = \frac{T_{fix}(x, y, z, t) - T_{fix}(x, y, z, 0)}{\overline{T_{fix}(t)} - \overline{T_{fix}(0)}} = \frac{T'_{fix}(x, y, z, t)}{\overline{T'_{fix}(t)}}, \quad (6.3)$$

where the double bar indicates a spatial average over the basin and the anomalies at time t are defined with respect to the initial state at time $t = 0$ (i.e. the pre-industrial control values). Similar definitions and normalisations are used for T'^n_{free} , $C^n_{ant(fix)}$, and $C^n_{ant(free)}$ (note that the preindustrial value of C_{ant} is zero everywhere, i.e. $C_{ant}(x, y, z, 0) = 0$). The normalisation ensures that the magnitude of the anomalies can be compared on the same scale.

Figure 6.6a-d shows the zonally integrated normalised anomalies for temperature and anthropogenic carbon in the fixed- and free-current experiments after 50 years (the same plots could be shown at different times, but 50 years was chosen because the patterns are clearest). There are noticeable differences between the patterns of T'^n_{fix} and that of T'^n_{free} , owing to circulation adjustments throughout the basin (Fig. 6.6e). However the patterns of anthropogenic carbon are similar for both the free and fixed currents experiments, apart from small differences in the mid-latitude surface, where heating causes a reduction in carbon solubility, and in the North Atlantic (Fig. 6.6f). T'^n_{fix} and $C^n_{ant(fix)}$ have similar patterns, except in the Southern Ocean AABW cell, with the presence of a carbon signal but no temperature anomaly. In the MITgcm Southern warming experiments ‘ $+T_S$ ’, and the $2 \times \text{CO}_2$ and $4 \times \text{CO}_2$ HadGEM2-ES experiments, there is a warming signal in the AABW. The discrepancy is therefore likely to be due to the choice in temperature restoring at the Southern boundary, which is kept at 0°C , therefore preventing the subduction of

heat. T_{fix}^n is expected to be more similar to $C_{ant(fix)}^n$ in the Southern Ocean than found in the MITgcm simulations, though fixed dynamics experiments with Southern boundary warming or a longer temperature restoring timescale are needed to verify this.

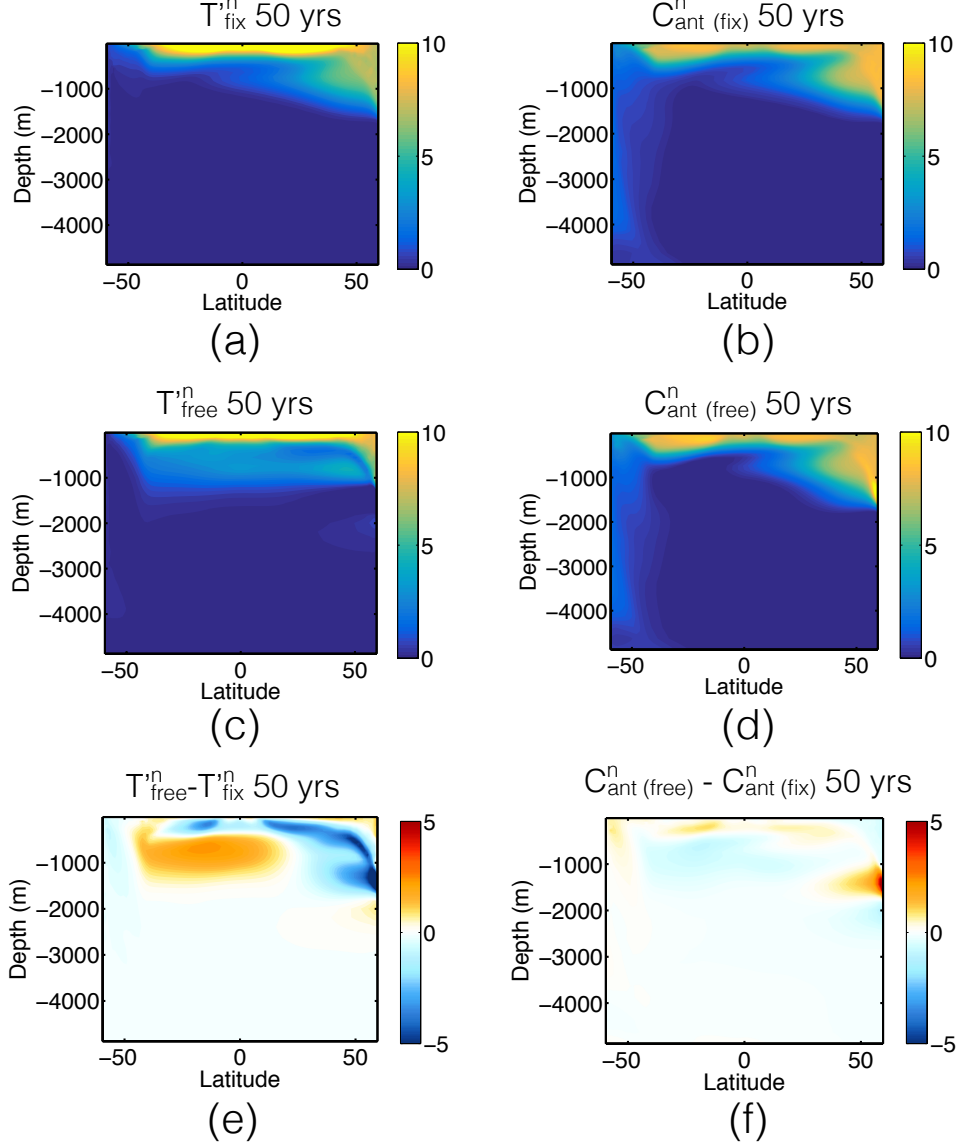


Figure 6.6: Zonal mean anomalies after 50 years in a) T_{fix} , b) $C_{ant(fix)}$, c) T_{free} and d) $C_{ant(free)}$, as well as the difference between fixed and free dynamics anomalies for e) temperature and f) anthropogenic carbon after 50 years in the MITgcm $1.1 \times T$ base warming experiment with 260ppm anthropogenic pulse emission. All anomalies have been scaled with respect to the global mean anomaly.

The similarity between the fixed and free C_{ant}^n patterns is partly due to anthropogenic carbon behaving like passive tracer [Khatiwala et al., 2009] but predominantly due to the small impact of warming-induced circulation on anthropogenic carbon uptake via the

compensation between saturated and disequilibrium effects (see section 5.2). Other studies such as Winton et al. [2013] also observe similar patterns in anthropogenic carbon uptake in free and fixed current experiments in a global Earth System Model. The patterns of $C_{ant(free)}^n$ are therefore similar to T_{fix}^n . To calculate the circulation-induced temperature anomaly field $T_{dynamics}^n$, we would need the T_{fix}^n and T_{free}^n fields, however, in the real ocean T_{fix}^n cannot be directly measured. Instead T_{fix}^n can be approximated as $T_{fix}^n \approx C_{ant(free)}^n$ leading to

$$T'_{fix}(x, y, z, t) \approx \overline{T'_{fix}(t)} C_{ant(free)}^n(x, y, z, t). \quad (6.4)$$

Equation 6.4 is a good approximation when considering zonally and/or vertically integrated quantities such as column ocean heat content H (see section 7.9 for detailed comparison plots).² However, since $T'_{fix}(x, y, z, t)$ cannot be directly measured in the ocean, neither can $\overline{T'_{fix}(t)}$. However, it is worth noting that estimating $C_{ant(free)}^n(x, y, z, t)$ is also difficult to measure in the real ocean due to the uncertainty in the methods to separate the anthropogenic carbon anomalies from the natural carbon. In section 6.1.1, it was shown that the fixed dynamics heat content is 28% larger than the free-evolving heat content under all warming scenarios up to timescales of 400 years. The difference in heat uptake can be used to make the approximation $\overline{T'_{fix}(t)} = 1.28 \overline{T'_{free}(t)}$. Therefore the pattern and magnitude of $T'_{fix}(x, y, z, t)$ can be reconstructed using

$$T'_{fix}(x, y, z, t) \approx 1.28 \overline{T'_{free}(t)} C_{ant(free)}^n(x, y, z, t). \quad (6.5)$$

Figures 6.7a-b (as well as figures in section 7.9 in the appendix) show the pattern of the reconstruction for the column-integrated heat content anomaly H'_{fix} , which matches the actual anomaly reasonably well (Fig. 6.7c), and that the magnitude of the anomaly is much improved with the 1.28-calibration factor (Fig. 6.7b). The error in the reconstruction is generally less than 20% outside the Southern Ocean for timescales of less than 200-300 years, after which the errors increase to 30% (likely due to the different history of the warming and carbon emissions in the MITgcm simulations).

Estimating the fixed temperature anomaly from the anthropogenic carbon signal is useful because the patterns of warming induced by the changes in the circulation can be estimated without the need to run fixed circulation experiments. The calibration factor is only needed for the magnitude of the anomaly. By separating temperature anomalies into

² $H = \int C_p \rho_0 T dz$ where C_p is the specific heat capacity and ρ_0 is a reference density. The vertical integral is evaluated over the full depth of the basin.

T'_{fix} and $T'_{dynamics}$, we can assess where the warming is directly driven by surface heat fluxes and where it is driven by dynamical adjustments (such as wind-induced changes in surface currents or adjustments of the large scale circulation due to buoyancy fluxes). To illustrate this, we apply eqn. 6.5 to the HadGEM2-ES $2\times\text{CO}_2$ experiment to calculate the vertically-integrated heat content change due to the dynamical adjustments $H'_{dynamics}$ (Fig. 6.8a) after 140 years.

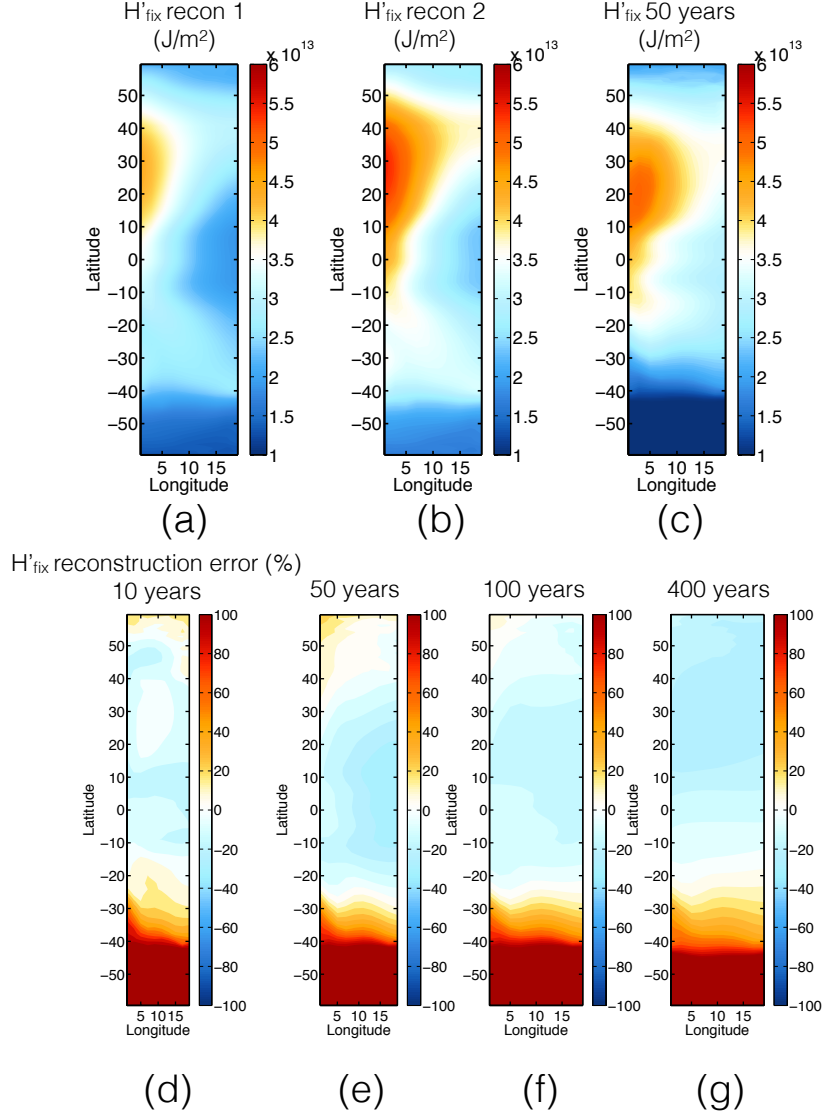


Figure 6.7: Fixed-current column integrated heat content anomaly H'_{fix} after 50 years from a) reconstruction using eqn. 6.5 without the calibration factor of 1.28 between $\overline{T'_{free}(t)}$ and $\overline{T'_{fix}(t)}$; b) reconstruction using eqn. 6.5 with the calibration factor of 1.28; c) directly from the fixed current 1.1xT MITgcm experiment. Panels d-g shows the percentage error in the reconstruction of H'_{fix} using eqn. 6.5 after d) 10, e) 50, f) 100 and g) 400 years.

The patterns of H'_{fix} in HadGEM2-ES (Fig. 6.8a) show that surface forcing warms the global ocean, particularly the North Atlantic and the Southern Hemisphere mid-latitudes. The pattern of $H'_{dynamics}$ (Fig. 6.8b) shows that circulation changes act to cool the mid-latitudes, shifting heat anomalies to the high latitudes in the North Pacific and particularly in the Southern Ocean where heat is more efficiently stored in the deep ocean. A similar effect is modelled in the fixed circulation experiments of Banks and Gregory [2006] and Winton et al. [2013] which gives us confidence that our method is valid. $H'_{dynamics}$ also shows cooling in the North Atlantic. In HadGEM2-ES, sea-ice in the North Pacific only covers the western side of the sub-polar gyre so the warming observed there is likely due to changes in gyre circulation. In the Southern Ocean, sea ice around Antarctica retreats between 90° and 150°E and between 200° and 300°E , such that part of the heat storage is related to ice retreat but a large part of the pattern is likely attributed to dynamically-induced circulation changes. In the North Atlantic, the slowing down and shallowing of the MOC is the dominant cause of the cooling pattern.

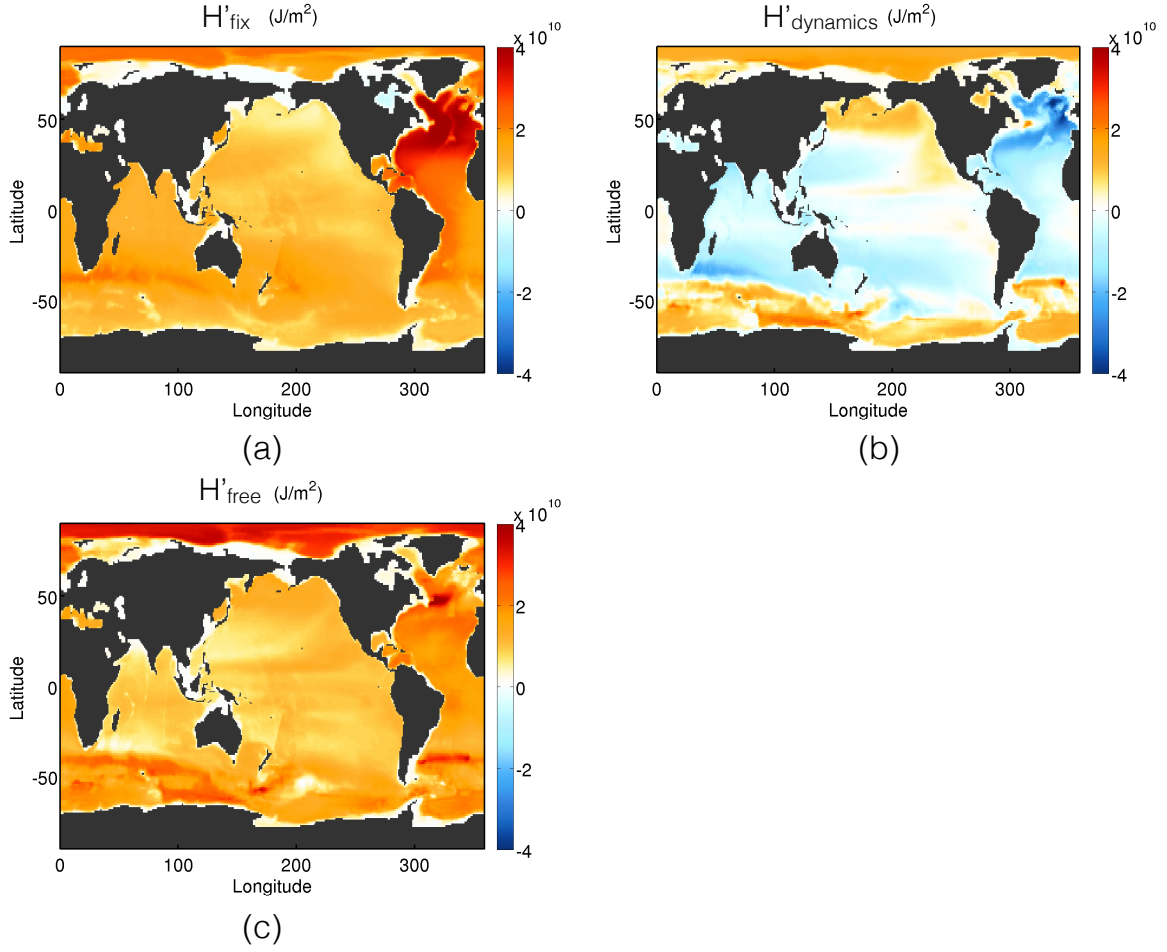


Figure 6.8: Vertically-integrated heat content anomaly due to a) surface warming and fixed pre-industrial circulation H'_{fix} , b) warming-induced dynamical changes $H'_{dynamics}$, c) freely evolving circulation H'_{free} calculated using eqn. 6.5 in the HadGEM2-ES $2\times\text{CO}_2$ experiment.

We can compare the patterns of the dynamically induced heat content anomaly with the free current anomaly H'_{free} (Fig. 6.8c). The warming patterns in the Pacific and particularly in the Indian Ocean show the mid-latitude cooling effect of the circulation anomalies on the overall warming pattern. In these areas, the circulation induced heat anomalies $H'_{dynamics}$ compensate a large part of the surface forced warming H'_{fix} . The free current heat anomaly H'_{free} also shows heat is stored in both the North Atlantic and the Southern Ocean. However, H'_{fix} shows that the majority of the heat due to the surface forced warming is stored in the North Atlantic. Therefore, the circulation anomalies act to shift a large part of the warming from the North Atlantic to the Southern Ocean, as seen in the patterns of $H'_{dynamics}$. The shift in heat storage could have implications for the timescales of ocean heat uptake and efficiency [Xie and Vallis, 2012, Winton et al., 2013].

Deducing the patterns of dynamically-induced warming is therefore useful for analysing the physical reasons behind the distribution of heat uptake. Particularly because local warming and cooling can affect local sea levels, precipitation patterns, weather and long term climate. However, further research is needed to fully understand the implications and physical mechanisms of heat re-distribution.

$H'_{dynamics}$ and H'_{fix} could be calculated for the real ocean using observations of temperature and carbon (although there are larger uncertainties in C_{ant}) to better understand the patterns of warming on short timescales, when the forcing history of heat and carbon is the same. The decomposition of T' into $T'_{dynamics}$ and T'_{fix} could therefore be used to calculate heat anomalies in the real ocean. As illustrated with the MITgcm wind stress and fixed dynamics temperature perturbation experiments, the net effect of dynamical changes on ocean heat content is significant. However, further study is required to quantify the accuracy of the procedure in the Southern Ocean especially.

Passive tracers have been used to understand ocean transport pathways in the past [e.g. Ewart and Bendiner, 1981, Sundermeyer and Price, 1998]. However classical passive tracers are advected along the perturbed circulation field, while anthropogenic carbon follows the mean state circulation more closely than classical passive tracers do, due to the carbon compensation. Passive tracer experiment in the real ocean also rely on the active release of a tracers so are therefore costly and limit the area of study to short timeframes and small areas. Global scale tracers such as oceanic chlorofluorocarbons (CFCs) provide global coverage and allow us to calculate the age of water parcels [e.g. Sonnerup, 2001, Dutay et al., 2002], though these require specific measurements of CFCs and suffer from errors such as long air-sea equilibration timescales. Additionally, the use of CFCs extends the timeframe of study to the introduction of CFCs in the 1940's. Our proposed method using anthropogenic carbon provides a potential method to estimate $H'_{dynamics}$ since the onset of anthropogenic warming in the pre-industrial era using only contemporary measurements, provided we can measure C_{ant} accurately and can estimate $\overline{T'_{free}(t)}$. Our method therefore extends the timeframe even further. Ocean carbon concentrations already form part of our observational system so the additional cost to estimate $H'_{dynamics}$ in the real ocean would be small.

Our proposed method is therefore a relatively simple new solution for providing long term insights into spatial patterns of ocean heat anomalies without the need of expensive model simulations, except to estimate the calibration factor between $\overline{T'_{fix}}$ and $\overline{T'_{free}}$. We expect the calibration factor to be a function of the ocean’s mean state circulation and surface warming pattern. Further research is required to determine whether the calibration factor can therefore be calculated using mean state quantities and surface warming patterns, without the need to run fixed dynamics experiments. However, the calibration factor only affects the magnitude of $H'_{dynamics}$. The patterns of the circulation-induced heat uptake already give us valuable insight into the physical adjustment of the ocean to climate change-induced warming.

6.2 Summary of chapters

The goal of this thesis is to provide a physical understanding of the ocean carbon cycle response to future climate change. From among the multitude of forcings that arise from anthropogenic carbon emissions, the focus was chosen to be on wind stress and surface warming since both of these play key roles in the large scale ocean circulation, both at depth and at the surface. At the outset of this thesis, we identified three specific goals: a) provide a theoretical understanding of the sensitivity of atmospheric CO_2 to high latitude wind stress and global ocean warming, b) disentangle the regional contributions of temperature and wind stress feedbacks on atmospheric carbon, and c) quantify the similarities of ocean heat and carbon uptake in a changing climate.

To address these goals, in chapter 3, an equation is derived for Southern Ocean $p\text{CO}_2$ as a function of wind stress. Southern Ocean $p\text{CO}_2$ increases with the area average wind stress according to a square root relationship with the driving physical mechanism being vertical mixing at the bottom of the mixed layer, while other contributions such as Ekman transport and biology vary little with wind stress. The sensitivity of the Southern Ocean $p\text{CO}_2$ response to wind stress is set by mixed layer DIC concentrations, mixed layer depth and vertical temperature gradient $\Delta_z T$. As atmospheric anthropogenic $p\text{CO}_2$ levels increase and surface temperatures changes, shifts in $\Delta_z T$ will determine the sensitivity of the Southern Ocean $p\text{CO}_2$ to wind stress.

In chapter 4, it is shown that increased Southern Ocean wind-stress strengthens the export of nutrients North of the Southern Ocean by increasing the nutrient stream. As a result, area-mean equatorial biological productivity increases. Due to stoichiometric ratios in the carbonate chemistry, the increased biological export from the surface to depth increases the Revelle Buffer Factor R in the Eastern tropical sub-thermocline shadow zone. Via the interior transport underneath the thermocline, the waters enriched in R are entrained into the nutrient stream and advected to the Northern sub-polar gyre, increasing the sub-polar gyre buffer factor and ocean $p\text{CO}_2$, and increasing atmospheric $p\text{CO}_2$. The non-local Southern Ocean wind stress feedback on atmospheric $p\text{CO}_2$ magnifies the local feedback by 80% to 110%. The non-local Southern Ocean feedback is twice as strong as the local North Atlantic wind stress feedback. The new results show that Southern Ocean winds have a much larger effect on the climate than previously thought. The novel mechanism is also relevant for any physical change affecting equatorial biology, not simply Southern Ocean wind stress and demonstrates a new concept: how an increase in biology can result in an increase in atmospheric $p\text{CO}_2$.

In chapter 5, we show that climate-change induced surface warming results in an increase in atmospheric $p\text{CO}_2$: over 400 years, 75% of the atmospheric $p\text{CO}_2$ anomaly is due to solubility and 25% due to biology. Dynamical changes associated with spatial patterns of warming have little impact due to dynamical carbon compensation between disequilibrium and saturated carbon. While atmospheric CO_2 anomalies are set by global warming, high latitude dynamics is key in the compensation mechanism, since we show that increased eddy activity in the Southern Ocean can break the carbon compensation. Over 3000 years, high latitude surface warming affects the overturning which in turn affects the balance of regenerated and saturated carbon. Over 3000 years, there is a compensation between saturated and regenerated carbon. The compensation increases with polar warming, resulting in small atmospheric CO_2 changes of about 10 ppm for 10°C of surface warming increase, in the strongest cases of Northern high latitude warming.

Finally, we use HadGEM2-ES and MITgcm simulations to show that ocean carbon transport is predominantly affected by circulation anomalies, while heat transport is affected by both temperature and circulation anomalies. We can exploit the similarities of the passive tracer component of temperature anomalies and anthropogenic carbon to calculate the extent to which circulation changes affect the ocean heat storage.

Caveats for each specific set of experiments are addressed in the respective chapters. However, across all chapters the coarse resolution of the MITgcm set-up is an important limitation, especially in the Southern Ocean which is key in the wind stress feedbacks as well as the surface warming experiments where it determines the compensation of saturated and equilibrium carbon on centennial timescales. While we believe the mechanisms for all results are still valid, the magnitude of the atmospheric $p\text{CO}_2$ may be reduced since the presence of eddies has been shown to dampen changes to the ocean’s density structure and large scale circulation [Munday et al., 2013, Jochum and Eden, 2015]. Similarly, eddies determine the timescale of adjustment of the pycnocline [Allison et al., 2011], so the timescales of the climate-carbon feedbacks may be affected. While we have attempted to quantify and illustrate the effects of eddies in chapters 3 and 5, additional simulations with an eddy resolving model are required to fully quantify the impact of eddies.

The size of the sector model is also a limitation which could have important implications. Firstly, the 20° degree width of the basin could affect the timescales of feedbacks. This is particularly relevant for the non-local Southern Ocean wind stress feedback, which depends on basin-wide advection. To address the concern, an identical MITgcm set up 40° wide is used to quantify the effects of basin width. It is found that the timescale of the non-local Southern Ocean wind stress feedback is longer in the 20° model and will therefore be longer for realistic geometries, though the magnitude of the feedback is unaffected.

Secondly, the 20° wide model results in a relatively weak overturning circulation (4.2 Sv), such that a reduction of 2 Sv results in a large fractional changes in the overturning. On the other hand, a decrease of 2 Sv in the overturning is a much smaller fractional change in the real ocean. However, in chapter 5 we show that the relevant quantity for atmospheric CO_2 feedbacks is f/Ψ , the ratio of vertical mixing over the overturning strength and not the absolute strength of the overturning. Lastly, the simple idealised geometry of the model places strong constraints of the circulation of the system. For example, there is no Pacific Ocean which could affect global carbon fluxes and large scale circulation. There is also no ocean South of the Southern channel, while in the real ocean, deep water is generally formed in the oceans South of Drake Passage. However, to fully understand the effects of the idealised geometry and size of the set-up, a realistic ocean bathymetry model is required, such as the set-up of Lauderdale et al. [2013].

In terms of the initial goals of the thesis, we believe the studies of Southern Ocean wind stress and global surface warming provide insight into the physical mechanisms of atmospheric CO₂ feedbacks. For example, using the scaling derived in chapter 3, we show that vertical mixing dominates the Southern Ocean wind stress feedback while the surface radiative flux is the key mechanism setting North Atlantic $p\text{CO}_2$ in response to wind stress perturbations. Moreover, the derived scaling can be used to predict atmospheric $p\text{CO}_2$ anomalies in the future without the need for expensive simulations, knowing only a handful of parameters from the control state.

The equilibrium climate feedback framework presented in section 2.4.2 provides a way of quantifying and decomposing non-local effects through the use of the non-local parameter ϵ_f which we apply to the wind stress studies to disentangle the specific regional contributions of wind stress feedbacks on atmospheric carbon. In the temperature study, we find that regional contributions to atmospheric CO₂ feedbacks lie in high latitude warming, since polar warming can neutralize a large part of the atmospheric CO₂ feedback due to solubility over 3000 years. Lastly, using anthropogenic carbon to calculate the dynamically-induced temperature anomalies in the ocean illustrates how circulation anomalies are the cause of the main differences in ocean heat and carbon uptake in a changing climate. Our work therefore shows the competition between dynamics and thermodynamics in setting the ocean heat storage and the climate response to warming.

6.3 Discussion and future work

This thesis quantifies the climate carbon feedback of wind stress and surface warming. To summarise the findings, we can apply our results to projections of the 21st century using CMIP5. Figure. 6.9 shows the atmospheric $p\text{CO}_2$ anomalies as a function of atmospheric CO₂-equivalent forcing with respect to the year 2000. The atmospheric CO₂-equivalent forcing is taken as the atmospheric $p\text{CO}_2$ anomaly at the year 2100 in each of the Recommended Concentration Pathways, RCP2.6 (70 ppm CO₂-eq. forcing), RCP4.5 (180 ppm), RCP6.0 (390 ppm) and RCP8.5 (1600 ppm) from the IPCC AR5.

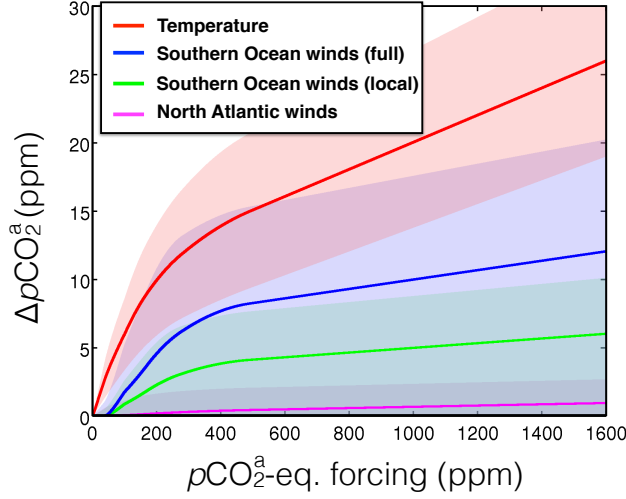


Figure 6.9: Atmospheric ΔpCO_2^a anomaly due to global mean sea surface temperature (red), full Southern Ocean wind stress (blue), local Southern Ocean wind stress (green) and North Atlantic wind stress (magenta) anomalies as a function of atmospheric CO_2 -eq. forcing at year 2100 derived from IPCC AR5 RCP scenarios, assuming the climate equilibrates for 400 years at year 2100 conditions.

The CMIP5 multi-model ensemble of global mean sea surface temperature change ΔT_G [Solomon, 2007], Southern Ocean wind stress anomaly, $\Delta \overline{\tau_{SO}}$ [Swart and Fyfe, 2012], and North Atlantic wind stress anomaly, $\Delta \overline{\tau_{NA}}$ [Gillett and Fyfe, 2013], are then taken at the year 2100 relative to the year 2000 for each RCP scenario. Assuming the climate equilibrates for 400 years at 2100 conditions,³ we use the MITgcm results and scaling equations to estimate the corresponding climate feedback to each forcing: ΔT_G (red), $\Delta \overline{\tau_{SO}}$ (blue: eqn. 4.10, green: eqn. 3.11) and $\Delta \overline{\tau_{NA}}$ (magenta: eqn. 4.15). The Southern Ocean wind stress feedback is then split into the full feedback which includes both the local and non-local effects (blue) and the local Southern Ocean feedback only (green). The confidence intervals are estimated using both the CMIP5 multi-model spread in ΔT_G , $\Delta \overline{\tau_{SO}}$ and $\Delta \overline{\tau_{NA}}$, uncertainty in the MITgcm response to ΔT_G and the range of atmospheric pCO_2 sensitivity to $\Delta \overline{\tau_{SO}}$ in the models tested in this thesis.

Figure 6.9 shows that given model projections for the year 2100, the temperature feedback is likely to be the strongest while the effects of North Atlantic wind stress are negligible. Uncertainty in the temperature feedback is also the smallest due to the carbon compensation over 400 years that limits the spread in atmospheric pCO_2 anomalies so uncertainty arises mainly from the projections of global mean temperature itself. If we

³We make this assumption simply to illustrate the relative strengths of the feedbacks.

chose to include the climate feedbacks after 3000 years of stabilization, the upper bound of the temperature feedback would remain the same while the lower bound would lie along $\Delta pCO_2^a = 0$ ppm owing to uncertainties in patterns of ocean warming. Uncertainty in the Southern Ocean wind stress feedback is large owing to the large range of atmospheric pCO_2 sensitivities to wind stress in CMIP5 models. However, the Southern Ocean wind stress feedback could be significant and even comparable to the temperature feedback, due to the contribution of the non-local effects. While the overall atmospheric pCO_2 feedback anomalies may seem small, a feedback of 10 ppm is equivalent to almost 10 years of anthropogenic emissions at the start of the 21st century. Furthermore, a change in atmospheric pCO_2 of 10 ppm could have significant impacts on warming, weather patterns and local ecosystems, thereby having further implications for climate change [e.g. Corti et al., 1999, Walther et al., 2002, Cao et al., 2005, Hinzman et al., 2005, Kelly and Goulden, 2008].

The derived scaling between atmospheric pCO_2 and Southern Ocean wind stress was derived with predicting future atmospheric CO_2 feedbacks in mind. However, the scaling can be applied to past states of the climate as well. Since the scaling is derived for a system in equilibrium, estimating 20th century atmospheric CO_2 anomalies is difficult. However, the scaling can be adapted to quantify the contribution of the Southern Ocean towards overall atmospheric pCO_2 levels in the pre-industrial climate, provided we have reasonable estimates of the stratification. In particular, there has been a strong focus on Southern Ocean winds and vertical mixing in explaining glacial-interglacial cycles of atmospheric CO_2 . Francois et al. [1997] and Anderson et al. [2009] suggest that there was an increase in vertical mixing in the Southern Ocean at end of the Last Glacial Maximum (LGM), 20000-12000 years before present, possibly related to an increase in wind stress. Using the scaling derived in chapter 3, we could therefore calculate the atmospheric pCO_2 related to those changes in vertical mixing. Francois et al. [1997] suggests that vertical mixing was increased by 20% to 50% at the end of the LMG, which would correspond to an increase in atmospheric pCO_2 of 15-35 ppm using our scaling.

So far, current theories of physical mechanisms have not been able to fully explain the glacial-interglacial cycles of atmospheric CO_2 [Sigman and Boyle, 2000, Solomon, 2007]. Marine biology is often used to try and explain the deficit in atmospheric CO_2 variations [Kohfeld et al., 2005]. The change in vertical stratification during the LGM discussed in

Francois et al. [1997] and Anderson et al. [2009] would induce a non-local atmospheric feedback in the North Atlantic of the type discussed in chapter 4, which has previously been ignored. Therefore, including the effects of the non-local atmospheric feedback in the North Atlantic due to Southern Ocean vertical mixing could aid our understanding of glacial-interglacial atmospheric CO₂ cycles.

Studies such as Anderson et al. [2002] have investigated changes in biological export in paleo climate, but only at a global scale. Early studies such as Broecker [1982], Pedersen [1983] and Farrell et al. [1995] study the effect of equatorial biological production in the Pacific on glacial-interglacial atmospheric CO₂ changes, though equatorial biology in the Atlantic has been left relatively unaddressed. Kohfeld et al. [2005] show that there was higher biological carbon export in the Atlantic equatorial shadow zone in the LGM but states that “these areas are relatively small and do not affect the overall picture,” while our research shows these areas should not be overlooked. Although global changes in ocean alkalinity have been acknowledged as being potentially important, there has been no consideration of spatial patterns. Kohfeld et al. [2005] does not give exact figures of the relative change in biological export in the Tropics and North Atlantic, so we cannot give an estimate of the atmospheric feedback due to Tropical biology. Therefore, further study using both past observations and model simulation with LGM conditions is needed to fully estimate the magnitude of the feedback, but we expect there to be a signal, given the results presented in this thesis.

While previous studies have been concerned with quantifying the atmospheric $p\text{CO}_2$ feedback of the biological pump using global mean efficiency of the biological pump [e.g. Ito and Follows, 2005], the non-local Southern Ocean wind stress feedback emphasises the importance of spatial patterns. When it comes to climate feedbacks, the focus has generally been on global means of temperature and atmospheric CO₂, as illustrated by the simple 1-D model in chapter 2. Perhaps more focus should lie on the spatial patterns, since Armour et al. [2013] shows that the time evolution of global feedbacks is tied to regional feedbacks. Wind-driven outgassing of natural carbon in the Southern Ocean will affect the atmospheric distribution of CO₂ and therefore spatial patterns of climate-concentration feedbacks (such as the carbon fertilisation effect [Kimball, 1983, Long et al., 2006]) and radiative forcing patterns.

Furthermore, the equation for Southern Ocean $p\text{CO}_2^o$ derived in chapter 3 is used to calculate an area average change in $p\text{CO}_2^o$. However, it could be used in the form of eqn. 3.5, without a basin-scale spatial average, to calculate spatial patterns of ocean $p\text{CO}_2$ anomalies. For example, we could use eqn. 3.5 to show that the sensitivity of $p\text{CO}_2^o$ to wind stress varies with longitude in the Southern Ocean, such that the $p\text{CO}_2^o$ in some areas is more susceptible to wind induced mixing than others. However, wind-induced vertical mixing does not only alter mixed layer DIC but also other mixed layer properties such as temperature. Wind-induced sea surface cooling will further affect regional feedbacks (as shown in Fig. 3.4). Therefore, a mixed layer flux equation as a function of wind stress could be derived for other climate-relevant quantities such as mixed layer temperature or nutrient concentration. By doing so, we could gain further insight into the effect of surface wind stress on regional climate feedbacks.

Similarly, the circulation-induced ocean heat content anomaly H'_{dynamics} derived in chapter 6 can be used to study regional feedbacks and the physical mechanisms that drive them. For example, the non-local Southern Ocean wind stress feedback arises from the inherent connection between the Southern Ocean and the North Atlantic sub-polar region, harking back to the ideas of Gnanadesikan [1999]. Similar principles could apply to ocean heat uptake, where changes in Southern Ocean dynamics could have impacts on ocean heat content in the North Atlantic.

Overall, this thesis provides a physical basis and quantification of the climate-carbon feedbacks of high latitude wind stress and global surface warming on atmospheric CO_2 levels. However, our current understanding of climate feedbacks as a whole requires much more research. We believe the work presented here advances our understanding of climate feedbacks and can be used as a valuable tool towards that goal.

Appendix

7.1 1-D Model re-optimization

The 1-D model used in section 2.1 is tuned by applying a likelihood profiling procedure [Pawitan, 2001] to determine five parameters: the timescale of deep ocean carbon advection, b_0 , surface heat and carbon uptake, a_2 and b_2 respectively, airborne fraction of CO₂ a year after emission, b_4 , and temperature feedback on natural carbon emissions, b_5 . It is useful to point out that the temperature feedback on emissions represents the response of the ocean carbon cycle to changes in temperature and can be either negative (an increase in ocean carbon uptake) or positive (net release of carbon from the ocean to the atmosphere) as a result of changes in dynamics, chemistry or biology.

To update the observational constraints, we use data from 1950 to 2012 for effective ocean heat capacity, 20th century greenhouse gas attributable warming and net airborne fraction of CO₂. The contribution of temperature feedback to airborne fraction of CO₂ and rate constant for advection of CO₂ into the deep ocean (Fig. 7.1d-e, effective timescale of surface to deep ocean transfer of carbon) are identical to Allen et al. [2009] and based on the C⁴MIP collection of models which feature full carbon cycles and we believed updating these numbers with CMIP5 data would not meaningfully affect the results. The values of the five optimised parameters are allowed to vary from the most likely value, after which the total likelihood of the projection is assessed and maximised by comparison with the optimisation constraints. Figure 7.1 shows the likelihood profiles, with the most likely value of the constraint with a likelihood of 1, for the old and the updated constraints. The net CO₂ airborne fraction (Fig. 7.1c) uses a longer time series. The greenhouse gas attributable warming (Fig. 7.1a) uses data from Gillett et al. [2013]. The effective ocean heat capacity (Fig. 7.1b) now includes the heat content change of the ocean over the period 1959-2011 [Lyman and Johnson, 2013, Levitus et al., 2012], which results in a large change in the effective heat capacity. An important cause of the change is the recent slowdown of global mean atmospheric temperature increase, which suggests a higher uptake of heat by the ocean. In addition, the best estimate of the fraction of greenhouse

gas (GHG) warming due to CO_2 has been revised downward from 0.7 to 0.68 (with a 0.58-0.98 95% confidence interval) by ensuring that the simple model's relationship between the 1861-2005 greenhouse gas-induced warming trend and Transient Climate Response (TCR) matches the suite of CMIP5 data from Gillett et al. [2013] and Forster et al. [2013]. The TCR of the simple model is varied by adjusting parameters in the energy balance equation after which the fraction of greenhouse gas warming due to CO_2 is adjusted until the 1861-2005 greenhouse gas-induced warming trend matches the CMIP5 mean. The constraints for the feedback contribution of temperature on natural emissions (and so on airborne fraction) and advection time constant of carbon transport from the surface to deep ocean (Fig. 7.1d and e) are kept as determined in Allen et al. [2009].

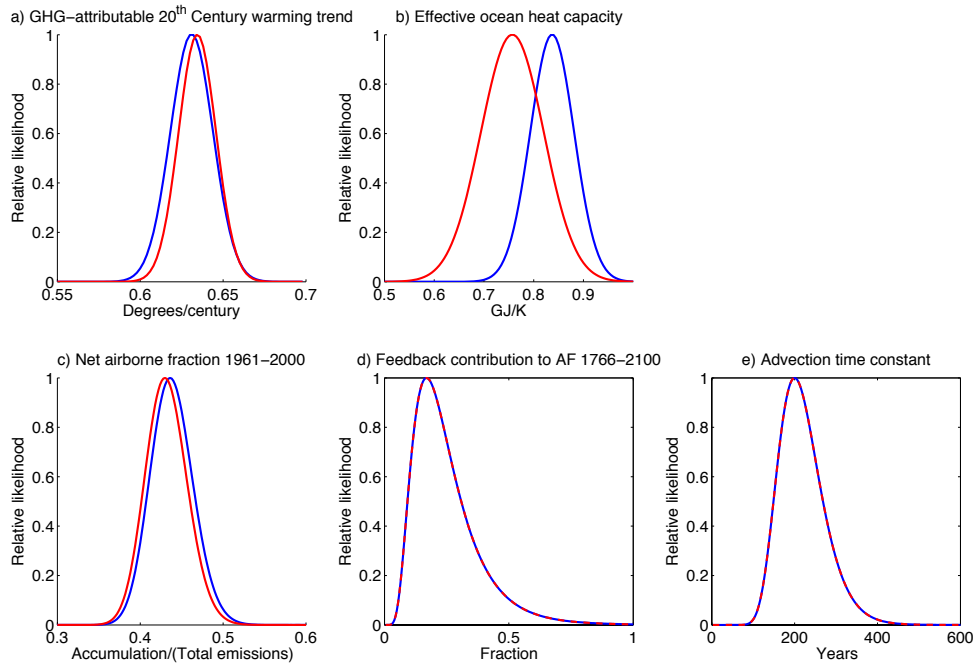


Figure 7.1: Likelihood profiles used in Allen et al. [2009] (red) and for updated constraints using recent data up to the year 2012 (blue) to optimise the box-diffusion model. The peak of the profiles indicate the most likely value of the constraint. The dashed lines indicate that the constraint remains unchanged.

Figure 7.2 shows global mean temperature anomaly projections and associated 66% confidence intervals (i.e. the IPCC “likely” range) using the old and new constraints. The most recent constraints result in a downward revision of peak warming to 1.3°C under a 1 trillion tonne of carbon emissions profile, compared to 2°C in Allen et al. [2009], along with a narrowing of the associated 66% confidence interval. The TCR is also reduced from 2.3°C to 1.4°C . The change in peak temperature and sensitivity is due to

the revised fraction of CO₂ induced GHG warming and the revised estimate of ocean heat capacity, which each contribute equally to the change. The Transient Climate Response to cumulative Emissions of carbon (TCRE) is the TCR divided by the cumulative emission of carbon at the time of atmospheric CO₂ doubling and gives a good indication of warming vs. emissions. In this case, the TCRE is 0.8 °C per TtC which is near the lower end of CMIP5 models [Matthews et al., 2009].

The timing of the peak warming is also delayed by 30 years for the same forcing. The delay can be explained by a combination of a stronger feedback of temperature on (natural) carbon emissions and the increased estimate of effective ocean heat capacity. The peak occurs in the year 2193, which is 172 years after emissions have begun to decrease.

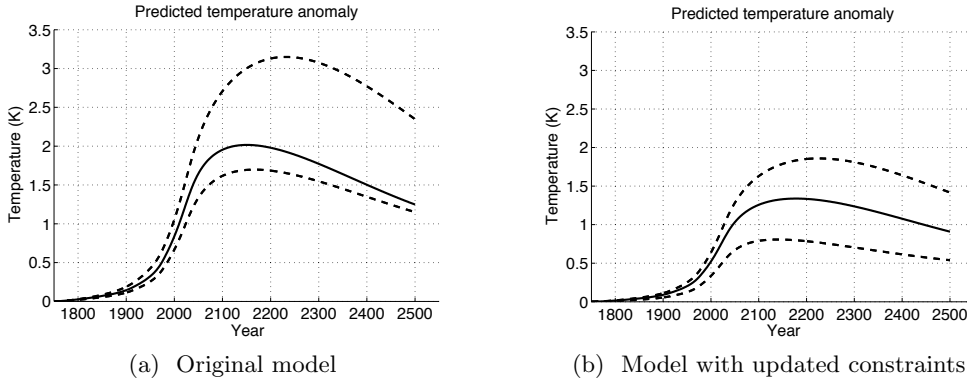


Figure 7.2: Global mean temperature anomalies for the most likely scenario (solid line) and 66% confidence intervals (dashed lines) for a) the box-diffusion model with Allen et al. [2009] constraints and b) the box-diffusion model constrained with observations up to 2012. Both models are driven by the same 1 trillion tonne of carbon cumulative emission profile.

7.2 MITgcm equations and parameters

We describe below the MITgcm model equations and basic parameters used in sections 3-6. The grid system employed in MITgcm can describe cartesian, spherical-polar or curvilinear coordinates with an Arakawa C-grid [Arakawa and Lamb, 1977]. Tracer quantities (e.g. temperature, salinity, etc.) are averaged over the grid cells while the zonal, meridional and vertical velocity components u , v and w respectively are defined at the centre of the orthogonal grid faces (Fig. 7.3). The velocity components in each grid cell are to be calculated as the average of the components on the bounding grid faces.

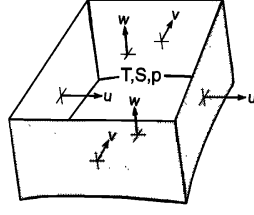


Figure 7.3: MITgcm grid cell: the velocity components u , v and w are defined at the centre of the orthogonal grid faces and tracer quantities are averaged over the volume of the cell.

The momentum equations solved in the model are the following:

$$\frac{\partial \bar{\mathbf{u}}_{\mathbf{h}}}{\partial t} + (\bar{\mathbf{u}} \cdot \nabla) \bar{\mathbf{u}}_{\mathbf{h}} + \mathbf{f} \times \bar{\mathbf{u}}_{\mathbf{h}} + \frac{1}{\rho_0} \nabla p = \nu_h \nabla^2 \bar{\mathbf{u}}_{\mathbf{h}} + \mathcal{F}_{\bar{\mathbf{u}}_{\mathbf{h}}} \quad (7.1)$$

with the 3D Eulerian velocity given by $\bar{\mathbf{u}} = (\bar{\mathbf{u}}_{\mathbf{h}}, w)$, $\bar{\mathbf{u}}_{\mathbf{h}}$ is the 2D horizontal Eulerian velocity vector, ∇ is the 2D vector gradient $\nabla = \hat{\mathbf{i}} \frac{\partial}{\partial x} + \hat{\mathbf{j}} \frac{\partial}{\partial y}$, ∇ is the 3D vector gradient, $\mathbf{f}$ is the Coriolis vector, ν_h is the 2D kinematic viscosity and $\mathcal{F}_{\bar{\mathbf{u}}_{\mathbf{h}}}$ is the forcing. The forcing $\mathcal{F}_{\bar{\mathbf{u}}_{\mathbf{h}}}$ is given by the surface wind stress vector $\boldsymbol{\tau}$:

$$\mathcal{F}_{\bar{\mathbf{u}}_{\mathbf{h}}} = \frac{\boldsymbol{\tau}}{\rho_0 \Delta_z z}, \quad (7.2)$$

where ρ_0 is a reference density and $\Delta_z z$ is the thickness of the surface layer. The pressure p is calculated using the hydrostatic approximation. The vertical velocity w is then solved using continuity:

$$\nabla \cdot \bar{\mathbf{u}}_{\mathbf{h}} + \frac{\partial w}{\partial z} = 0. \quad (7.3)$$

Lastly, a free surface is used so that $z = \eta$ at the surface, where η is the sea surface height defined as the deviation from $z = 0$, so that η can be positive or negative. Potential temperature (θ) and salinity (S) follow the tracer equations:

$$\frac{\partial \theta}{\partial t} + \mathbf{u} \cdot \nabla \theta = \nabla \cdot (\boldsymbol{\kappa} \cdot \nabla \theta) + \mathcal{F}_{\theta}, \quad (7.4)$$

$$\frac{\partial S}{\partial t} + \mathbf{u} \cdot \nabla S = \nabla \cdot (\boldsymbol{\kappa} \cdot \nabla S) + \mathcal{F}_S, \quad (7.5)$$

where $\boldsymbol{\kappa}$ is the 3D vertical diffusion matrix, and $\mathcal{F}_{\theta}$ and $\mathcal{F}_S$ are source/sink terms. $\mathbf{u}$ is the residual velocity field $\mathbf{u} = \bar{\mathbf{u}} + \mathbf{u}^* = (\bar{\mathbf{u}}_{\mathbf{h}} + \mathbf{u}_{\mathbf{h}}^*, \bar{w} + w^*)$, where the terms $\mathbf{u}^* = (\mathbf{u}_{\mathbf{h}}^*, w^*)$ represent the bolus velocity due to unresolved eddies. The effect of unresolved eddies is parametrized using the Gent-McWilliams parametrization [Gent and McWilliams, 1990, GM90] via the skew-flux formulation of Griffies [1998], which incorporates diapycnal diffusion by rotation of the ‘vertical’ diffusivity [Redi, 1982]. The horizontal bolus velocity

Variable	Symbol	Value	Units
Harmonic viscosity		0.0075	m^4s^{-1}
GM diffusivity	κ_{GM}	1000	m^2s^{-1}
Diapycnal diffusivity		3×10^{-5}	m^2s^{-1}
Restoring timescale T	λ_θ	30	days
Restoring timescale S	λ_S	10	days
Vertical viscosity		10^{-3}	m^2s^{-1}
Time step (physical model)		900	s
Time step (biogeochemical model)		43200	s
Redfield ratio carbon to phosphorous	$r_{C:P}$	117	
Redfield ratio nitrogen to phosphorous	$r_{N:P}$	16	
Fraction of new production going to DOP	γ	0.67	
Phosphate half saturation constant	K_N	5×10^{-4}	mol m^{-3}
Light half saturation constant	I_k	30	W m^{-2}
DOP remineralisation timescales	λ_{remin}	6.43×10^{-8}	s^{-1}

Table 7.1: Model Parameters for the MITgcm at $2^\circ \times 2^\circ$ horizontal resolution.

$\mathbf{u}_h^*$ is calculated from the vertical derivative of the bolus streamfunction Ψ_{GM} , which is equal to the isopycnal slope in the x- and y-direction $\mathbf{S} = (S_x, S_y, 0)$ multiplied with the GM thickness diffusivity κ_{GM} such that

$$\mathbf{u}_h^* = -\frac{\partial}{\partial z} \Psi_{GM} = \frac{\partial}{\partial z} [\kappa_{GM} \mathbf{S}]. \quad (7.6)$$

The value of κ_{GM} depends on the model resolution and Ψ_{GM} is set to zero at the upper and lower boundary. Since S_x is effectively infinite in the mixed layer, Ψ_{GM} is tapered in the mixed layer. Horizontal diffusion of tracers is incorporated in the Griffies [1998] skew-flux formulation of the GM90 parametrization. Potential temperature at the surface is forced with a climatological relaxation to monthly values θ^c according to a timescale λ_θ similar to Munday et al. [2013]. Similarly, salinity is forced at the surface with a climatological relaxation towards monthly values S^c with a timescale λ_S [Haney, 1971]. The forcing are given by

$$\mathcal{F}_\theta = -\lambda_\theta(\theta - \theta^c), \quad (7.7)$$

$$\mathcal{F}_S = -\lambda_S(S - S^c). \quad (7.8)$$

Such a restoring boundary, as used in other studies [e.g., von der Heydt and Dijkstra, 2008], assumes that the ocean is forced by a steady climatological atmosphere [Haney, 1971], which is only true at equilibrium without variability in the forcing, i.e. when θ^c and S^c are constants in time (aside from the climatological seasonal cycle). The values of constants used in the model are summarized in table 7.1.

7.3 Mixed layer carbon equation

This section describes the integration of the mixed layer carbon tracer equation over the vertical extent of the mixed layer section 2.4.1. The mixed layer carbon flux equation is given as:

$$\frac{\partial C}{\partial t} = -\nabla \cdot [(\bar{\mathbf{u}} + \mathbf{u}^*)C] + \nabla \cdot (\boldsymbol{\kappa} \cdot \nabla C) + F_{air-sea} + F_{bio}, \quad (7.9)$$

where $\bar{\mathbf{u}}$ is the Eulerian velocity, $\mathbf{u}^*$ is the bolus velocity and $\bar{\mathbf{u}} + \mathbf{u}^*$ is the residual velocity [Marshall and Radko, 2003], $\boldsymbol{\kappa}$ is the diffusivity matrix, $F_{air-sea}$ is the air-sea flux and F_{bio} the biological flux, as defined in section 2.3.1. The vertical integration limits of the mixed layer tracer equation for carbon are the free surface $\eta(x, y, t)$ and the mixed layer depth $H(x, y, t)$. Therefore, we need to apply Leibniz' rule to simplify the integrals. For the derivative with respect to t :

$$\int_{-H}^{\eta} \frac{\partial C}{\partial t} dz = \frac{\partial}{\partial t} \int_{-H}^{\eta} C dz - C(\eta) \frac{\partial \eta}{\partial t} - C(-H) \frac{\partial H}{\partial t} = (H + \eta) \frac{\partial C_h}{\partial t}. \quad (7.10)$$

It also assumed that $C(-H) = C(\eta) = C_h$. For the derivative with respect to horizontal spatial coordinates, with the horizontal velocity vector $\mathbf{u}_h = (\bar{u} + u^*)\hat{i} + (\bar{v} + v^*)\hat{j} = u\hat{i} + v\hat{j}$:

$$\int_{-H}^{\eta} \nabla \cdot (C \mathbf{u}_h) dz = \nabla \cdot \int_{-H}^{\eta} C \mathbf{u}_h dz - C_h \mathbf{u}_h \cdot \nabla \eta - C_h \mathbf{u}_h \cdot \nabla (-H). \quad (7.11)$$

We've assumed $C(x, y, z) = C_h(x, y)$ which is constant with z , so we can write the integral:

$$\nabla \cdot \int_{-H}^{\eta} C \mathbf{u}_h dz = \nabla \cdot \left(C \int_{-H}^{\eta} \mathbf{u}_h dz \right) = (\nabla \cdot C) \left(\int_{-H}^{\eta} \mathbf{u}_h dz \right) + C \frac{\partial h}{\partial t}, \quad (7.12)$$

where we have used the vertical integral of the continuity equation, such that:

$$\int_{-H}^{\eta} \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) dz = \frac{\partial h}{\partial t} + \nabla \cdot \int_{-H}^{\eta} (\mathbf{u}_h) dz = 0, \quad (7.13)$$

and the kinematic boundary condition, in the absence of freshwater forcing:

$$w = \frac{\partial h}{\partial t} + \mathbf{u}_h \cdot \nabla h \quad (7.14)$$

to substitute $\nabla \cdot \int_{-H}^{\eta} (\mathbf{u}_h) dz$ for $\frac{\partial h}{\partial t}$, where $h = \eta + H$ (see White et al. [2008]). Therefore the net result of the vertical integration of the horizontal advection term gives:

$$\int_{-H}^{\eta} \nabla \cdot (C \mathbf{u}_h) dz = (\nabla \cdot C) \left(\int_{-H}^{\eta} \mathbf{u}_h dz \right) + C \frac{\partial h}{\partial t} + \mathbf{u}_h \cdot \nabla h. \quad (7.15)$$

We can therefore use eqn. 7.15 to substitute for $\int_{-H}^{\eta} \nabla \cdot (C \mathbf{u}_h) dz$ in the mixed layer carbon eqn. 2.24 which leads to to full equation integrated over the mixed layer:

$$\begin{aligned} h \frac{\partial C_h}{\partial t} = & - C_h(w_h + u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y} + \frac{\partial h}{\partial t}) + \frac{\partial C}{\partial x} u h + \frac{\partial C}{\partial y} v h \\ & + \left[\kappa_x \frac{\partial^2 C_h}{\partial x^2} + \kappa_y \frac{\partial^2 C_h}{\partial y^2} \right]_h + \left(\kappa_z \frac{\partial C_h}{\partial z} \right)_{z=-H} \\ & + h F_{air-sea} + [F_{bio}]_h. \end{aligned} \quad (7.16)$$

7.4 Mixed layer entrainment velocity

We describe below the derivation for the ocean mixed layer entrainment velocity w_h used in section 3.1.2. For an ocean mixed layer of density ρ and depth h , a wind stress τ across the surface will act as a source of mechanical energy input (W). The rate of change of the energy W can be calculated by considering a force (wind stress) accelerating an object (mixed layer) with a velocity u_{ml} :

$$\frac{dW}{dt} = \tau u_{ml} = \rho_0 u_*^2 u_{ml} = m_0 \rho_0 u_*^3, \quad (7.17)$$

where u_* is the frictional wind speed and m_0 is a constant. The input of mechanical energy must be equal to the rate of change of potential energy PE :

$$\frac{dPE}{dt} = \frac{d}{dt} \left[\int_{-h}^0 \rho_0 g z dz \right] = \frac{1}{2} g h^2 \frac{d\rho_0}{dt} \quad (7.18)$$

Equating the rate of change of mechanical and potential energy then gives:

$$m_0 \rho_0 u_*^3 = \frac{1}{2} g h^2 \frac{d\rho_0}{dt} = \frac{1}{2} g h^2 \left(\frac{d\rho_0}{dT} \right) \left(\frac{dT}{dt} \right) \quad (7.19)$$

The term $\frac{d\rho_0}{dT}$ is equal to $-\alpha \rho_0$ where α is the thermal expansion coefficient. To calculate the term $\frac{dT}{dt}$ we considering the influence of a heating at the surface in the form of a net surface heat flux Q and a cooling at the bottom due to an entrainment of colder deep water:

$$\frac{dT}{dt} = \frac{Q}{\rho_0 h C_p} - \Delta_z T \frac{1}{h} \frac{dh}{dt} = \frac{Q}{\rho_0 h C_p} - \Delta_z T \frac{1}{h} w_e \quad (7.20)$$

where C_p is the specific heat capacity and the rate of change of the mixed layer depth is set equal to the entrainment velocity w_e . Using this expression in equation 7.19 allows us to solve for the mixed layer entrainment velocity:

$$w_h = \frac{2}{\alpha g h \Delta_z T} \left(m_0 u_*^3 - \alpha g h \frac{Q}{2 \rho_0 C_p} \right) \quad (7.21)$$

In essence, the entrainment velocity results from balancing in the kinetic energy due to wind friction with the change in potential energy, resulting from a change in density due to heating at the surface and mixing of cold water at the mixed layer base.

7.5 40° longitude wide basin MITgcm model set-up

The experimental set-up for the 40° longitude wide basin MITgcm model is identical to the 20° wide version described in section 2.3 except for the width of the basin. The equations

and parameters are the same as stated in section 7.2. Figure 7.4 shows the control mean circulation of the model and Fig. 7.5 shows control mean patterns of DIC, mixed layer biological productivity, ocean $p\text{CO}_2$ and air-sea CO_2 flux. The equilibrium atmospheric $p\text{CO}_2$ of the model is 295 ppm. The main difference between the 40° longitude and 20° longitude wide model is the strength of the MOC, 6.2 Sv and 4.2 Sv respectively. The wider basin also results in longer zonal advection timescales, since the time taken for water to be advected from the Eastern to the Western side of the basin is longer in the 40° longitude model.

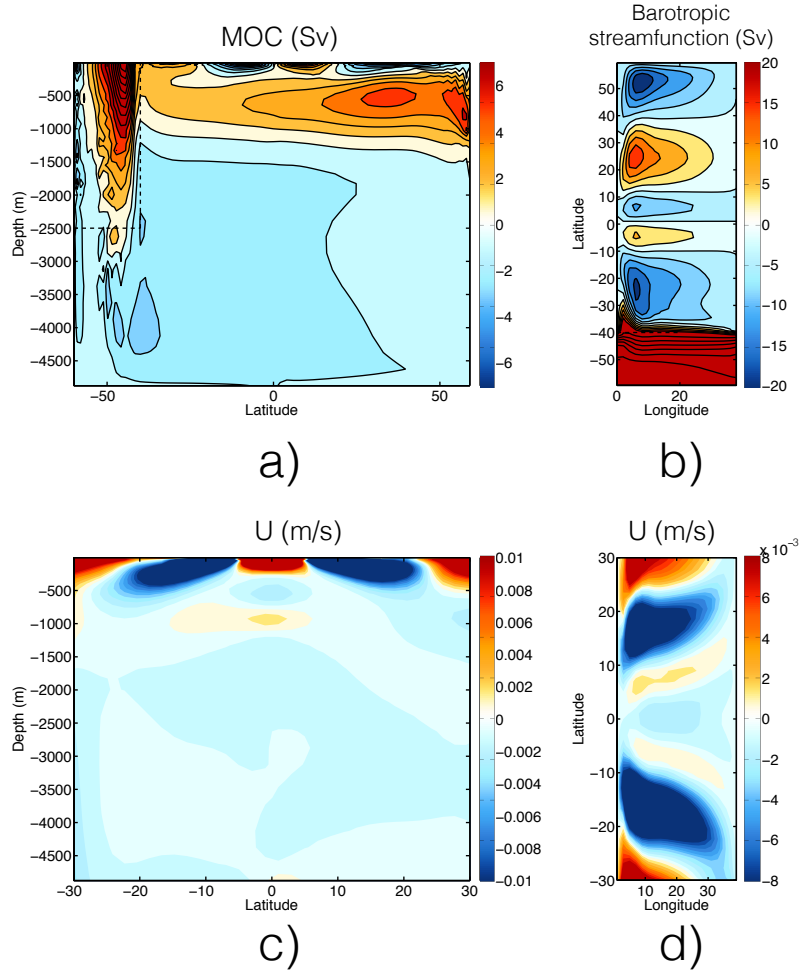


Figure 7.4: a) Control mean MOC, b) control mean barotropic stream function, c) residual zonal mean of the zonal velocity u as a function of depth and latitude and d) residual zonal velocity u at a depth of $z=-400\text{m}$ for the 40° wide basin MITgcm model.

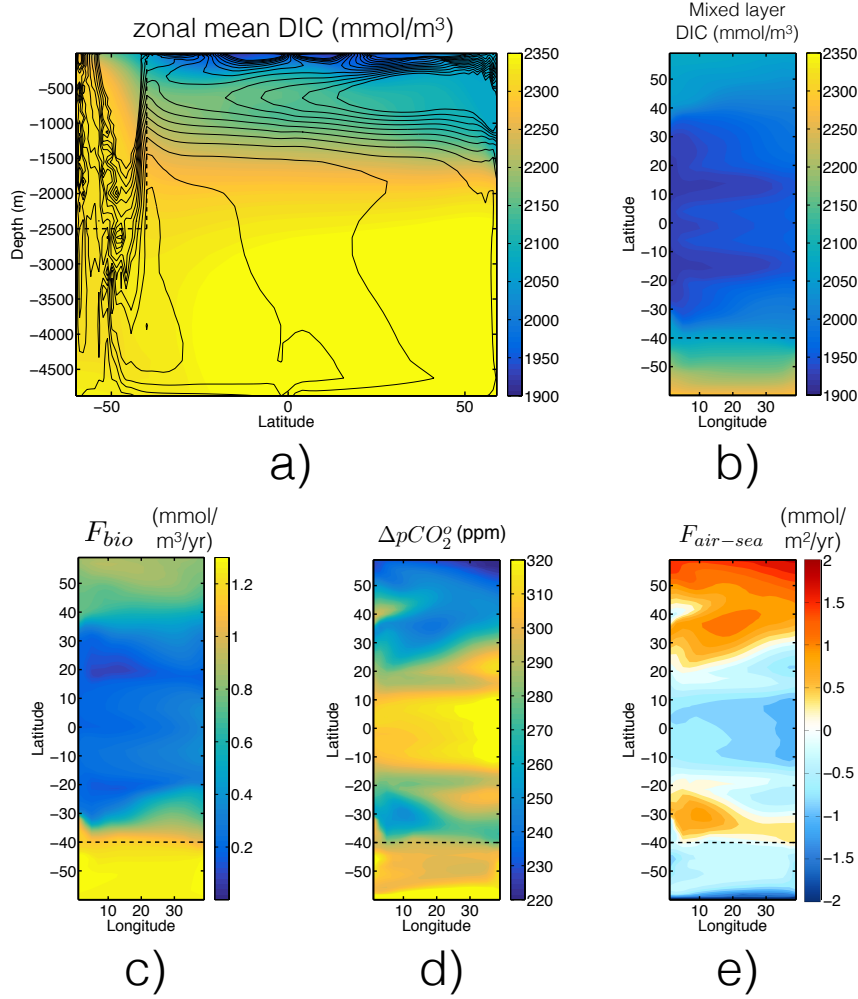


Figure 7.5: a) Zonal mean DIC, b) mixed layer DIC, c) mixed layer mean biological productivity, d) ocean surface $p\text{CO}_2$ and e) air-sea CO_2 flux for the 40° wide basin MITgcm model.

7.6 Southern Ocean wind stress perturbations for a 40° longitude wide basin and triangular wind stress profile

To verify that the relationship between Southern Ocean $p\text{CO}_2$ and area-mean wind stress $\overline{\tau_{SO}}$ derived in section 3.1.2, eqn. 3.7, is independent of the wind stress profile and basin size, two sets of experiments are run. Firstly, we present the results from a set of wind stress perturbation experiments where the sinusoidal wind stress profile between $60^\circ - 20^\circ\text{S}$ is replaced with a piecewise linear function with the same maximum value of $\tau = 0.15 \text{ N/m}^2$ at 40°S (Fig. 7.6a). The second set of wind stress perturbation experiments feature a doubled basin size, such that the sector is 40° wide instead of 20° (Fig. 7.6b). In each case, we use the scaling equation 3.7 to estimate Southern Ocean $p\text{CO}_2$ as a function of $\overline{\tau_{SO}}$ and find that the scaling holds regardless of basin size and wind profile (note that the

sensitivity of the relationship varies between the sets of experiments since the climatologies are slightly different).

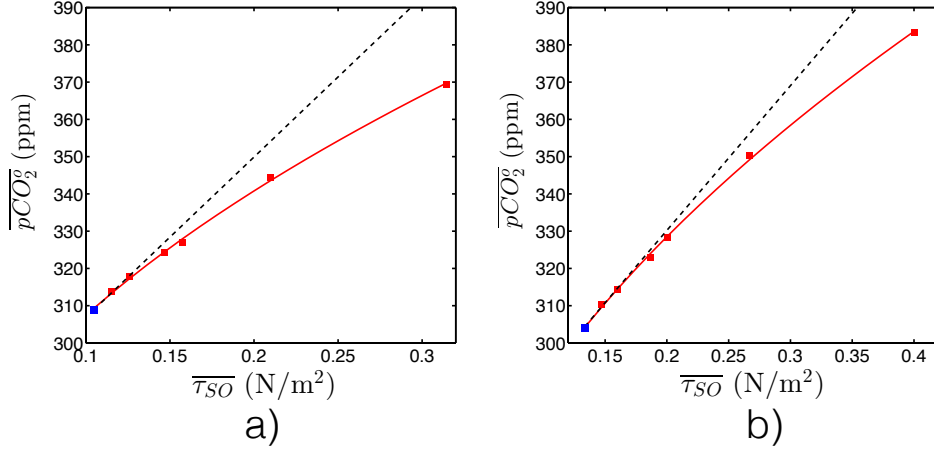


Figure 7.6: Response of the MITgcm Southern Ocean $\overline{pCO_2}$, as a function of mean channel wind stress. The blue square denotes the control experiment and the red squares the perturbed experiments (after 400 years of model run) for a) a piecewise linear wind stress profile between $60^\circ - 20^\circ S$ and b) for a model domain 40° wide. The red lines are the calculations made using eqn. 3.7 and the dashed black lines are linear trends around the control state.

7.7 Ecological models for primary production: nutrient and light limitation perturbation experiments

Below we describe the nutrient and limitation model used MITgcm. The nutrient and light limitation parameters in the model are perturbed to simulate changes in surface biological productivity without any dynamical changes, as used in section 4.1.4. The rate of change in DIC in the mixed layer is modelled in GCMs by considering a source-minus-sinks function, SMS for the primary production (usually phytoplankton population P):

$$SMS(P) = V_P(T) \gamma(I, N) P - sinks, \quad (7.22)$$

where $V_P(T)$ is the temperature-dependent maximum growth rate (which only some models include) and $\gamma(I, N)$ is a dimensionless function from 0 to 1 that gives the combined influence of nutrients N and irradiance I on the growth rate. The form of the function $\gamma(N)$ is empirically determined to take a hyperbolic functional form (Monod 1949):

$$\gamma(N) = \frac{N}{K_N + N}, \quad (7.23)$$

where K_N is the half-saturation constant (or Monod constant) which is the concentration of N at which point the growth rate is at half its maximum value. The value of K_N

is found to be generally less than 3 mmol m^{-3} [Eppley et al., 1969]; in the MITgcm experiments, the default value is 0.4 mmol m^{-3} . By perturbing this parameter in a model, the strength of the primary production at given concentration of N can therefore be increased or decreased. The change will be stronger in areas that are severely limited by macro-nutrients such as Nitrogen and Phosphorous (i.e. the Tropics and mid-latitude), while high-latitude areas that are light-limited will see a smaller percentage change.

The same principle applies for the function $\gamma_P(I)$, which is most accurately represented by the form [AD and Platt, 1976]:

$$\gamma_P(I) = \frac{I}{\sqrt{I_k^2 + I^2}} \quad (7.24)$$

where I_k is the irradiance half-saturation constant, which takes a default value of 30 Wm^{-2} in MITgcm. By perturbing the parameter I_k , the strength of biology can be perturbed for a constant value of local I and the change will be strongest in areas that highly light-limited (i.e. high latitude areas).

7.8 Calculating ΔpCO_2^a due to Southern Ocean alone and calculation of ϵ_f

To calculate the non-local feedback parameter ϵ_f in the Southern Ocean wind stress perturbation experiments in section 4.1.4, we use the equation:

$$\Delta pCO_2^a = (1 + \epsilon_f \frac{P_E}{P_{SO}}) P_{SO} \Delta \overline{pCO_{2SO}^o} + \Delta P_{SO} \overline{pCO_{2SO}^o} + \Delta P_E \overline{pCO_{2E}^o}, \quad (7.25)$$

where the subscript SO refer to the Southern Ocean where the wind stress perturbation is applied and the subscript E refers to the rest of the ocean basin. P_{SO} and P_E are the area fractions of the respective regions, weighed by the local wind stress and Δ indicates the anomaly with respect to the control mean. We solve eqn. 7.25 for ϵ_f using values of ΔpCO_2^a and $\Delta \overline{pCO_{2SO}^o}$ from the MITgcm perturbation experiments. Here, we present a separate estimate of ϵ_f using a more rigorous approach for the case of Southern Ocean wind stress perturbations.

Firstly, we explicitly calculate the atmospheric pCO_2 change due to the perturbation in the Southern Ocean pCO_2 alone, $(\Delta pCO_2^a)_{SO}$, without any non-local effects. Secondly, we

calculate the full pCO_2^a including non-local contributions, $(\Delta pCO_2^a)_{NL}$. The two equations we solve are therefore:

$$(\Delta pCO_2^a)_{SO} = P_{SO} \Delta \overline{pCO_{2SO}^o} + \Delta P_{SO} \overline{pCO_{2SO}^o} + \Delta P_E \overline{pCO_{2E}^o} \quad (7.26)$$

$$(\Delta pCO_2^a)_{NL} = (1 + \epsilon_f \frac{P_E}{P_{SO}}) P_{SO} \Delta \overline{pCO_{2SO}^o} + \Delta P_{SO} \overline{pCO_{2SO}^o} + \Delta P_E \overline{pCO_{2E}^o} \quad (7.27)$$

By combining the two equations, we can then solve for ϵ_f :

$$\epsilon_f = \frac{(\Delta pCO_2^a)_{NL} - (\Delta pCO_2^a)_{SO}}{P_E \Delta \overline{pCO_{2SO}^o}}. \quad (7.28)$$

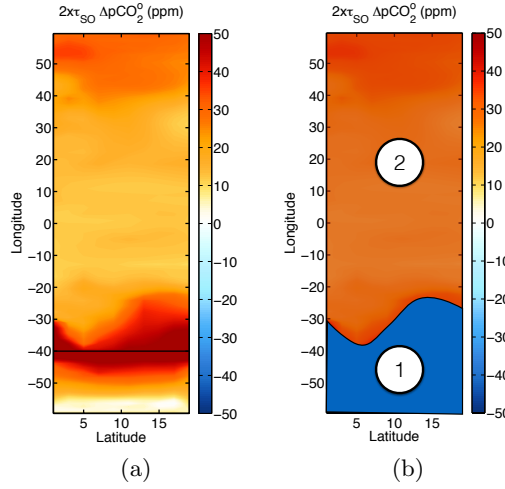


Figure 7.7: a) Change in ocean pCO_2 in the $2 \times \tau_{SO}$ experiment and b) the corresponding regions 1 and 2 used to calculate the local and non-local pCO_2^o contribution towards atmospheric pCO_2 .

To calculate $(pCO_2^a)_{SO}$, we explicitly integrate the equilibrium air-sea flux equation over the whole domain to solve for atmospheric pCO_2 :

$$\int F_{air-sea} dA = \int K \tau (pCO_2^a - pCO_2^o(f)) dA = 0. \quad (7.29)$$

$pCO_2^o(f)$ indicate perturbed values of pCO_2^o while $pCO_2^o(0)$ are the control values (see below). However, we use only time-varying pCO_2^o values from the perturbation experiments in the Southern Ocean where the wind stress perturbation is applied. We keep the pCO_2^o values elsewhere in the basin constant using control run values: the pCO_2^o from region 1 in Fig. 7.7 is time-varying, while the pCO_2^o from region 2 is kept constant (region 1 is calculated as the area in the Southern hemisphere with a higher than average pCO_2^o anomaly, including the Southern channel). We therefore solve the equation:

$$\int^1 K \tau (pCO_2^a - pCO_2^o(f)) dA + \int^2 K \tau (pCO_2^a - pCO_2^o(0)) dA = 0, \quad (7.30)$$

for pCO_2^a , where integration over region 1 is indicated by the subscript 1 on the integral sign, and similarly for region 2. τ kept at the unperturbed value for both region to ensure P_{SO} is constant.

For the full non-local pCO_2^a , we simply use the time-varying pCO_2^o values from the perturbation experiments for the whole basin.¹ We therefore solve the following equation for pCO_2 :

$$\int^1 K\tau(pCO_2^a - pCO_2^o(f))dA + \int^2 K\tau(pCO_2^a - pCO_2^o(f))dA = 0, \quad (7.31)$$

The result of the calculation is shown in Fig. 7.8. The values of ϵ_f shown in Fig. 7.8 are slightly higher than those shown in Fig. 4.9 for low wind stress perturbations. The order of magnitude is similar however and both calculations show ϵ_f increasing with Southern Ocean wind stress.

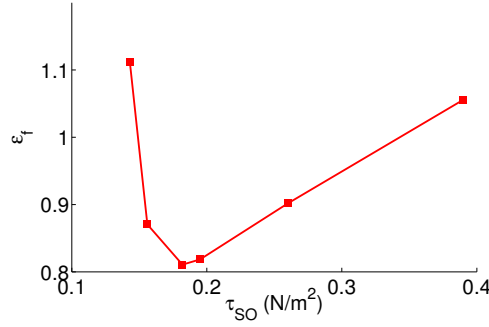


Figure 7.8: ϵ_f as a function of wind stress in MITgcm.

7.9 Temperature and carbon anomalies in fixed circulation experiments

In section 6.1.2, the similarity between the normalized fixed and free C_{ant}^n patterns is exploited to calculate the circulation-induced temperature anomaly field $T_{dynamics}^n$. The following figures show further examples of the reconstruction of the normalized circulation-induced temperature anomaly field $T_{dynamics}^n$. Using anthropogenic carbon anomalies in the freely evolving MITgcm experiments, $C_{ant(free)}^n$, we can calculate $T_{dynamics}^n$ by using the equation:

$$\begin{aligned} T_{dynamics}^n(x, y, z, t) &= T_{free}^n(x, y, z, t) - T_{fix}^n(x, y, z, t) \\ &\approx T_{free}^n(x, y, z, t) - C_{ant(free)}^n(x, y, z, t). \end{aligned} \quad (7.32)$$

¹The scaling in chapter 3 effectively estimates $(pCO_2^a)_{SO}$.

Figure 7.9 shows the reconstruction of the zonal mean $T_{dynamics}^n$ field against the modelled $T_{dynamics}^n$ field at year 10 (panel a), 50 (panel b), 100 (panel c) and 400 (panel d) in the 1.2xT MITgcm experiment. The modelled $T_{dynamics}^n$ field is calculated by taking the difference between the temperature anomalies between freely evolving and fixed dynamics temperature anomalies in the 1.2xT MITgcm experiments with an anthropogenic CO₂ pulse of 260 ppm.

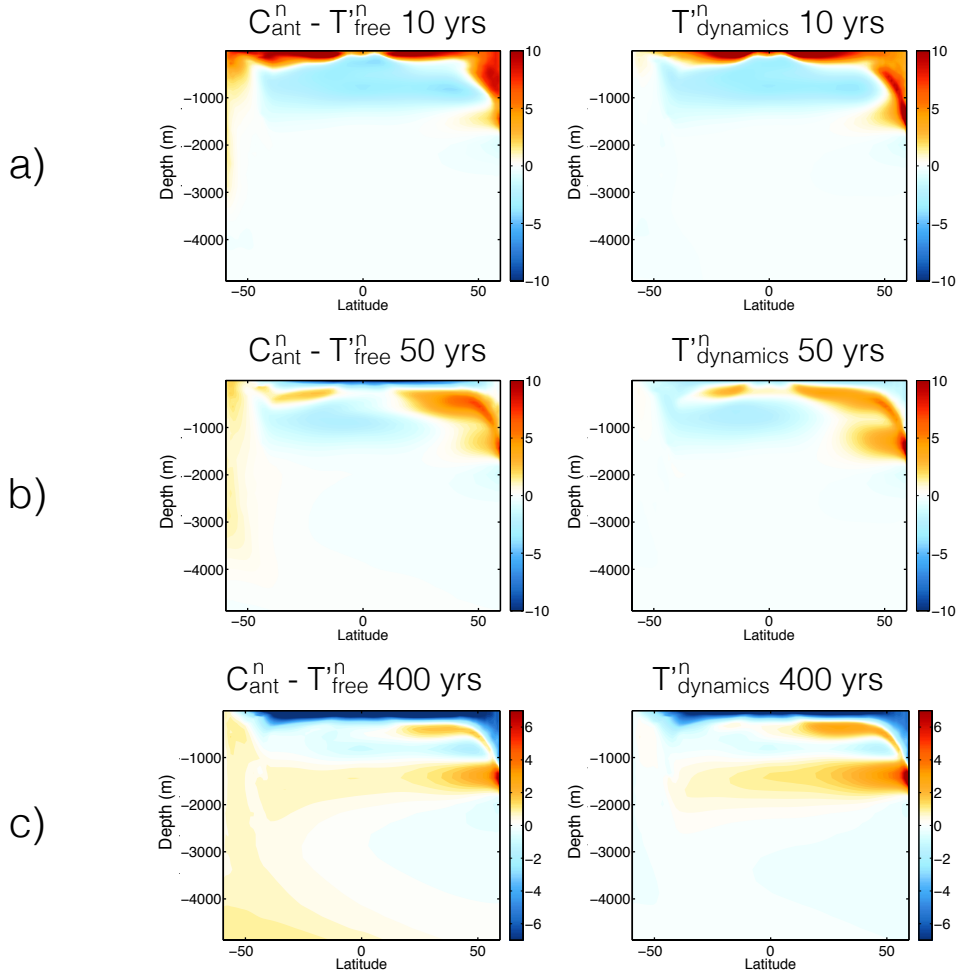


Figure 7.9: Zonal mean reconstruction (left) and model output (right) of $T_{dynamics}^n$ at a) 10, b) 50, c) 100 and d) 400 years in the MITgcm 1.1xT base warming experiment with 260ppm anthropogenic pulse emission.

The temperature anomalies can be integrated vertically to give the column integrated heat content H . Figure 7.10 then shows the reconstruction of $H_{dynamics}^n$ field against the modelled $H_{dynamics}^n$ field at year 10 (panel a), 50 (panel b), 100 (panel c) and 400 (panel d) in the 1.2xT MITgcm experiment. Figures 7.9 and 7.9 both show the same

features described in section 6.1.2. Firstly, we see discrepancies between the reconstructed dynamically-induced temperate anomalies and the modelled values in the Southern Ocean (likely due to the temperature restoring at the Southern boundary). Secondly, the error in the reconstruction increases with time and becomes significant after several centuries (see Fig. 7.9d in particular. The error is likely due to changes in the circulation affecting the anthropogenic carbon distribution over centennial timescales and the different forcing histories of temperature and carbon in the MITgcm simulations.

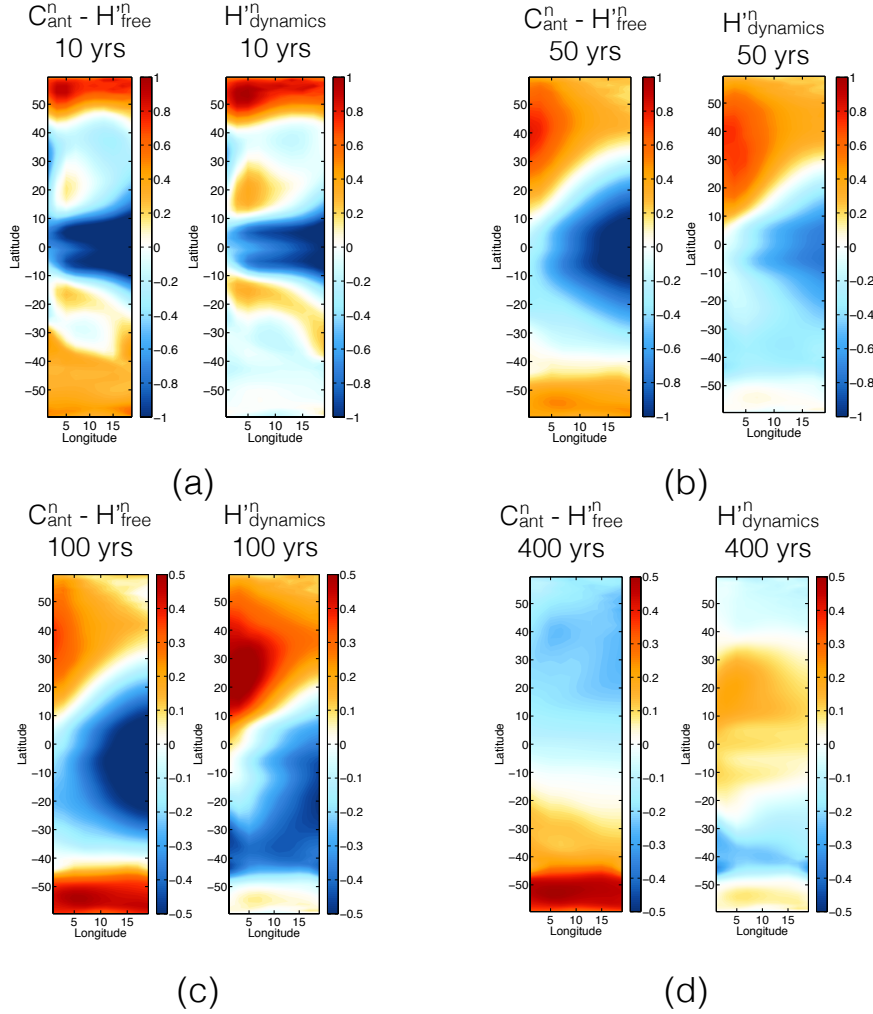


Figure 7.10: Reconstruction (left) and model output (right) of vertical column heat anomaly H_{dyn}^n at a) 10, b) 50, c) 100 and d) 400 years in the MITgcm 1.1xT base warming experiment with 260ppm anthropogenic pulse emission.

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