An Analysis of the Hydrological Cycle and Poleward Heat Transports Simulated by Two Climate Models

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Student Declaration

I hereby declare that the work presented in this thesis is my own and that all else is appropriately referenced.

Christopher Dancel,
April 2012
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Abstract

Heat and Freshwater transport by the oceans and atmosphere are an integral part of the climate system, aiming to cool the Tropics and warm the Extra-Tropics. General Circulation Models (GCMs) are used to simulate the climate system, however a key weakness to them is the uncertainty associated with model predictions. One component of this uncertainty is due to the model structural bias associated with the choice of ocean model vertical coordinate type, which can have substantial feedback within a coupled ocean-atmosphere model. This thesis aims to investigate the Heat and Freshwater transport in the climate system with specific relevance to three main topics:

1. sensitivity of heat and freshwater transport to model numerics
2. coupling between heat and freshwater transport
3. changes to heat and freshwater transport to increasing $CO_2$ concentration

Firstly, the choice of ocean vertical coordinate on the computed heat and freshwater transport in the oceans and atmosphere was investigated. By comparing the models CHIME (isopycnal level ocean model) and HadCM3 (z-level ocean model) in a control climate, it was found that variations to the atmospheric latent and dry static energy transports were much larger than those induced from anthropogenic emission scenarios predicted by the latest IPCC report (AR4, 2007).

Secondly, a new theory that constrained the ratio of ocean to atmospheric heat transport $H_o/H_a$ as a function of ocean temperature and salinity was examined. This theory was tested using a control scenario from the HadCM3 model for mid-latitudes, finding good agreement over the Northern Hemisphere, though poorer performance over the Southern Hemisphere.

Finally, climate snapshots in CHIME were analysed under an increasing $CO_2$ environment. An examination of heat and freshwater transport for the ocean - atmosphere and the atmospheric dry static - latent energy components, showed significant compensation within each pair. Further investigation into the ocean overturning circulation and atmospheric moisture transport revealed: a salinification (freshening) of the Atlantic (Pacific), increased zonal moisture transport through Central America, and a weakening of the Atlantic meridional overturning circulation, validating CHIME’s anthropogenic responses with predictions from AR4.
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CHAPTER 0

Introduction

In the fourth IPCC assessment report (AR4, Solomon et al., 2007) it was stated that “Warming of the climate system is unequivocal, as is now evident from observations of increases in global average air and ocean temperatures, widespread melting of snow and ice and rising global average sea level.”, based on the latest observational and model data at that time. More recently, the addition of 2011 to our historical time series record on global mean surface temperature has for the first time, made increases in global mean temperature statistically significant at the 95% confidence level, a common confidence boundary. Changes in the climate system have thus observed significant warming trends over the last few years, which have been popularised under the phrase ‘Global Warming’. The common consensus as to the origins of this warming have been pointed towards in the increase in greenhouse gases (GHGs) into the atmosphere since the start of the industrial revolution (AR4), a major one being $CO_2$. Measurements of atmospheric $CO_2$ by Keeling (1960) since 1958 have indeed revealed an increasing trend, expected from the combustion of fossil-fuel.

The link between increasing $CO_2$ and global warming is a contentious issue, with one camp preaching anthropogenic reasons (e.g. IPCC reports), whilst the other being inclined to natural variability of the climate (e.g. work by Willie Soon, Patrick Michaels and Richard Lindzen). Whatever one believes it is essential to understand the physical mechanisms of the climate so as to predict the worst case scenarios.

Predictions about the state of our climate system and how it will evolve in the future usually
involve general circulation models (GCMs). These couple the atmosphere, ocean, land and ice areas to produce a climate model that has credible observational similarities. Unfortunately the predictions of such models for the coming century differ widely. The GCM ensemble from the third assessment report (Houghton et al., 2001) were forced with a plausible emissions scenario (SRES IS92a) for the coming future. Though all models showed an increase in global mean surface temperature, some showed a substantial weakening of the meridional overturning circulation (MOC) of the Atlantic, others showed a slight weakening, and one model showed no significant change. This latter model was the ECHAM4/OPYC model (Roeckner et al., 1996) where its ocean model employed an isopycnic (constant density) vertical coordinate system, in comparison to the rest of the other models that used constant depth. Under the fourth assessment report (Solomon et al., 2007), a similar range in behaviours were exhibited though differences between model predictions were much less linked to ocean coordinate type (Megann et al., 2010).

In another case, Sun and Bleck (2001b) ran a coupled climate model that included the isopycnic ocean model HYCOM (Bleck, 2002). Even after running the model forced with increasing CO2 at 1% per year, no changes were observed to the strength of the MOC. However when run again with a different configuration of HYCOM (Sun and Bleck, 2005), the model showed a significant decrease in MOC strength. The connection therefore between changes to the climate state and vertical coordinate system is still unclear. Since the majority of climate models still use a constant depth configuration, it is important to include models that are structurally different so as to understand the architectural bias’ that arise from any model comparisons, but also as an extra check to verify any previous predictions.

By their design, isopycnic models eliminate spurious diapycnal mixing within its advection scheme resulting in certain advantages over constant depth level models. These include: better representations of near-adiabatic flows over sloping isopycncal surfaces (Megann and New, 2001) such as deep western boundary currents and equatorial under currents, the removal of spurious mixing of dense waters at sill overflows (Roberts et al., 1996), and the preservation of water properties over long time and length scales (Roberts et al., 1996; Megann et al., 2010). Its disadvantages include a reduction in vertical resolution for regions of weak stratification and vaguely defined detrainment from the mixed layer.

Apart from a model’s coordinate choice, the intrinsic mixing schemes of each are entirely unique. In particular for processes near the surface where heat is mixed into the interior, these differences can lead to feedbacks in the atmosphere. Megann et al. (2010) compared
the control climate simulations for the models HadCM3 and CHIME. Both models shared identical atmospheric models and were of similar ocean resolution; the primary difference was that HadCM3 utilised a constant z ocean vertical coordinate system with a corresponding Bulk Formula mixing scheme (Kara et al., 2000), whilst CHIME used a constant density (isopycnic) vertical coordinate with complimentary K-Profile Paraterisation (KPP) mixing scheme. When compared to climatological maps of SST, HadCM3 was found to have an overall cold SST bias, whilst CHIME had an overly warm SST bias particularly over the southern ocean. Megann attributed these changes to the lower summertime mixing properties of CHIME over the region that led to the trapping of heat at the surface. Variations in SST belonging to ocean model architectural setup will ultimately feed back into the entire climate system, firstly affecting the atmospheric circulation that eventually is felt by the ocean through its wind driven circulation. The details of model architectural bias are still currently being tackled and a large part of this thesis aims to understand the sensitivity of climate models to the choice of vertical coordinate system.

One of the important aspects of climate science is the ability to predict changes with resolutions suitable for local study rather than on a global emphasis. Suitable scales can range from an area the size of the UK to a whole continent (HadRM3H, Hulme et al., 2002). For this task regional climate models (RCMs) are employed that have resolutions of the order of 50km compared to that of a GCM of around 300km. RCMs are usually initialised with conditions at its model boundary using data outputted from a coarser GCM. For any regional climate prediction to have any credibility the GCM providing it data must have an accurate representation of the present climate in the first place.

The benchmarks of a ‘good’ climate model are those that reproduce the climate to a fair degree of accuracy. These may include realistic surface temperature distributions, the major ocean and atmospheric circulations, a close imitation of the hydrological cycle, and internal variability for events such as ENSO, to name a few examples. Only the most sophisticated GCM models that fulfilled this type of criteria were included in the IPCC AR4, with estimates in climate change being the most precise presently available.

An important constraint on the climate is known as the Clausius–Clapeyron (C-C) relationship, which states that for an air parcel of fixed relative humidity (an index of how much an air parcel is saturated), the specific humidity (how much moisture a parcel contains) will increase at a rate of approximately 7% per Kelvin warming (Held and Soden, 2006). From
Figure 1: Scatterplot of the percent change in global-mean, column integrated water vapour versus the percentage change in global-mean, surface air temperature for the PCMDI AR4 models under the SRES A1B forcing scenario. Differences are taken between the first twenty years of the 21st and 20th century. Figure taken from Held and Soden (2006).

the analysis of the model ensemble forced with possible GHG scenarios (SRES), a unanimous response across all models witnessed an increase in global mean surface temperatures with direct consequences of increases in surface evaporation and moisture content of the atmosphere. The response of the hydrological cycle to increased warming has been of particular interest. For the Tropical region (Held and Soden, 2006; Vecchi and Soden, 2006) it was found that the most robust responses of the climate included: the decrease in convective mass fluxes, the increase in horizontal moisture transport, the associated enhancement of the pattern of evaporation minus precipitation and its temporal variance, and the decrease in the horizontal sensible heat transport in the extratropics, all of which were the result of an increase in lower tropospheric water vapour content at the C-C rate (see Fig. 1). For the Extra-Tropics (Lorenz and DeWeaver, 2007), model ensemble responses showed that poleward moisture and precipitation rates increased at a rate less than that stated by the CC rate (see next chapter) with precipitation increasing the least. Precipitation changes were weakest during the warm periods when energy constraints are more important. It was also found that changes to evaporation was not well correlated with changes to temperature or zonal mean wind speed, but rather intermodel differences correlated better to changes in the oceans meridional heat transport. These responses presently act as model behavioural ‘benchmarks’ to be tested against other future models. As mentioned early, given that the majority of GCM models within the AR4 were primarily design with a constant depth coordinate system in its ocean model, it is interesting to note if these responses will be as robust when dealing with an isopycnic ocean model. The results will be important in two ways: firstly they act as a validation for the predictions made by previous models, and or secondly, they can shed light on any new climate mechanisms that
may not be modelled very well with constant depth coordinate models. Within this thesis, an examination of a new isopycnic coupled climate models aims to answer these questions.

A very important feature of our climate is the transport of heat; the net surplus of radiation entering at the Tropics and net deficit within the Extra-Tropics drives a poleward transport of heat via the oceans and the atmosphere, with little influence from the land (fig. 2). Different heating rates across all latitudes causes pressure gradients in the atmosphere that drive winds. This in turn feeds back into the ocean through surface wind stress which then drive surface currents and the majority of the ocean circulation (Wunsch and Ferrari, 2004).

At the same time, surface heating leads to evaporation of seawater that gets transported by the atmosphere and eventually falls as precipitation. Changes to the surface freshwater balance have mitigating influences on the density of seawater with consequences for the general circulation of the ocean. Within the atmosphere, the presence of water vapour (the major GHG) increases an air parcel’s ability to absorb surface radiation and the retention of energy. At the same time however water vapour leads to clouds that have a negative feedback on warming, with its high albedo property reflecting incoming solar radiation back into space. Heat and freshwater are therefore linked together by highly complex processes. Progress has been made to understand the limits of our climate system, constraining climate behaviour to a few key variables.

Figure 2: Required total heat transport from the TOA. These are compared to the Ocean Heat transport, and residual Atmospheric heat transport (RT - OT). Data is from NCEP Re-analysis. Figure taken from Trenberth and Caron (2001).

In the atmosphere, the transport of heat can be decomposed into parts belonging to that of moisture (latent) and dry air (dry static energy). Atmospheric heat and freshwater transport
are therefore coupled through the latent heat component (Pierrehumbert, 2002). Calculations of the latter can be made from the divergence patterns of evaporation and precipitation. At the ocean surface the vertical exchange of moisture leads to an imprint on the salinity \((S)\) of the ocean, with precipitation (evaporation) leading to a salinification (freshening) of local water. Whilst the examination of the transport of salinity can infer ocean freshwater transport, so can the transport of ‘temperature’ \((T)\) infer a heat transport. All water parcels in the ocean have a distinct temperature and salinity that when mapped in T-S space, generate a distribution for all water masses in a certain T-S range. This distribution is referred to as the T/S curve and has been used to couple heat and freshwater transport in the ocean (Stommel and Csanady, 1980).

Recently, new theory has emerged that has extended the work of Stommel and Csanady that couples the ratio of ocean to atmospheric heat transport simply from the T/S curve. The assumptions of baroclinicity for the atmosphere limit this theory to extra-tropical latitudes only. Previous work on this ratio for the same region was conducted by Czaja and Marshall (2006). By decomposing the heat transport of the ocean and atmosphere to parts belonging a mass streamfunction \(\Psi\) and a suitable temperature difference \(C\Delta\theta\), they found that the temperature difference between the two media were comparable, with the main difference being \(\Psi_A >> \Psi_O\), where the subscripts relate to the atmosphere and ocean. This new theory bypasses the need for any information on the circulation strength of the ocean or atmosphere, greatly simplifying the equations to describe our climate. If true, it offers a new mechanism to constrain heat transport behaviour simply by the temperature and salinity profiles of the ocean. The theory is not intended to be a better method of calculating the ratio ocean to atmospheric HT since satellites already provide superior data in that respect. The aim is to provide an alternative approach to the problem that will further stimulate debate on the present theories of heat transport. How accurate the theory holds will be tested within a coupled climate model.
0.1 Aims of this Thesis

This thesis covers a broad range of topics however the underlying trend throughout encompasses the analysis of *Heat and Freshwater Transport* within the climate system. During the research period of this thesis the author was under contract to carry out diagnostics on a new GCM, CHIME (Megann et al. 2010). Three months was spent at the National Oceanography Centre running in-house diagnostics and other research. Consequently, much of the model analysis within this thesis originates from the CHIME model. Three main areas of this topic are described and discussed, these being the:

1. sensitivity of heat and freshwater transport to model numerics (Chapters 3 and 4)
2. coupling between heat and freshwater transport (Chapter 5)
3. changes to heat and freshwater transport to increasing $CO_2$ concentration (Chapters 6 and 7)

Specifically we will answer the following questions:

**Question 1.**

*In a coupled model, how does changing from constant vertical depth to a constant vertical density coordinate system in the ocean affect a). meridional heat transport [Chapter 3]; b). hydrological cycle [Chapter 3]; and c). the representation of the ocean circulation [Chapter 4]?*

A sub-topic to this research explored early on was testing a theory that linked the ratio of ocean to atmospheric heat transport $H_o/a$ to temperature and salinity variations in the ocean. From this, we ask,

**Question 2.**

*How well does a new theory that predicts $H_o/H_a$ for extra-tropical latitudes hold when tested in a coupled climate model [Chapter 5]?

For the ‘end’ of the thesis we aim to answer questions pertaining to the robust responses of the climate system to a warming climate. This section additionally acts as a way to verify the responses of the climate already previously observed with other models. We ask,
Question 3
Within a warming climate, how will a). the structure of meridional heat transport change [Chapter 6], b). the freshwater transport change [Chapter 6], and c). the ocean circulation change [Chapter 7]?
CHAPTER 1

Earth’s Climate System

In this chapter we outline the basic features of the climate system, describing the ocean, atmosphere, land, and ice systems. The mathematical framework is first described, following onto the application of its use to explain such dynamics as the ocean and atmospheric general circulation patterns.

1.1 Earth’s Radiation Budget

At Earth’s present distance from the Sun, the amount of incoming solar radiation through a unit of area is $S_0 = 1368 \text{Wm}^{-2}$ where the latter quantity is referred to as the Solar Constant. Measurements of this value has shown variations of approximately 0.1-0.3% (Eddy et al., 1982), though for the sake of computational analysis, $S_0$ is taken as a constant.

If the Earth was a perfect black body the radiative balance equation between incoming solar radiation and outgoing terrestrial radiation would be

$$S_0(1 - \alpha)\pi R^2 = 4\pi R^2 \sigma T^4$$

where $\alpha$ is the albedo of the Earth’s surface, $R = 6371 \text{km}$ is the radius of the Earth, $\sigma = 5.67 \times 10^{-8} \text{Js}^{-1} \text{m}^{-2} \text{K}^{-4}$ is Stefan’s Constant, and $T$ is temperature in Kelvin. Re-arranging eq. 1.1 we get

$$T = \left[\frac{S_0(1 - \alpha)}{4\sigma}\right]^{1/4}$$

(1.2)
If this Earth was a perfect absorber i.e. $\alpha = 0$, this would equate to a global mean temperature of roughly 256K, or $-17^\circ C$, far colder than what is presently observed. The discrepancy between this value and a present global mean surface temperature value of approximately $15^\circ C$ is that the latter configuration does not include the effects of either the ocean or the atmosphere. Outgoing Terrestrial, or infrared radiation is absorbed by the atmosphere and re-radiated back onto the surface, raising surface temperatures to what they are now. The presence of clouds, land vegetation and snow cover does serve to increase the planetary albedo (especially snow cover), though this effect is of secondary importance when balanced against the warming effect that the ocean and atmosphere generate.

Considering a conceptual model of the Earth that included both the warming effects of the ocean and atmosphere, but was a completely rigid body, the difference in temperature between that found at the equator and at the poles would be far larger than what is presently observed. Again this is not a true observation in our present climate due to a characteristic found both in the ocean and atmosphere - they are fluid. What we mean by this is that they are free to move around through advection from place to place. The importance of this observation is that with this movement comes the transport of local properties throughout the system, for example, energy. Since there is more incoming solar radiation per unit area near the Tropics ($30^\circ S - 30^\circ N$) rather than near the poles, this meridional temperature gradient is greatly reduced by the advection of heat from low to high latitudes by the ocean and atmosphere. These latter two media are thus extremely important in mitigating both global mean and regional temperatures.

1.2 Energetics of the Climate System

Incoming solar radiation onto the Earth comes in the form of high energy shortwave (SW) radiation. The atmosphere is mostly transparent to this type of radiation with the majority being able to reach the surface of the Earth without much interference. As the surface starts to warm it emits its own form of low energy longwave (LW) radiation. The atmosphere is translucent to this type of radiation, with some being absorbed by the atmosphere and re-emitted towards the surface; the rest escapes into space as terrestrial radiation. With the former, this mechanism is the basis of ‘global warming’ with so called ‘greenhouse gases’ such as $H_2O$ and $CO_2$, re-absorbing LW radiation and leading to a large increase in global surface temperatures that are far outside of Earth’s natural variability (see section 1.5 below for details).
Incoming radiation at the Earth’s equator will have a greater power concentration per unit area than the same amount of radiation incident at high latitudes over a greater area. Incoming fluxes are thus weighted by the \( \cos(\text{latitude}) \) to take this effect into account (see Fig. 1.1).

We can characterise the net radiation balance at the top of the atmosphere (TOA) as

\[
R_{\text{TOA}} = SW_{\text{in}} - SW_{\text{out}} - LW_{\text{out}}
\]

(1.3)

where the subscripts ‘in’ and ‘out’ refer to the incoming and outgoing components of radiation incident on the Earth’s surface. The spatial distribution of these quantities are shown in Fig. 1.2 along with their zonal mean distributions in Fig. 1.3. These diagnostics (and similar figures to come) are taken from the model HadCM3 (see section 2.2 for details) and not from real world observations. The motivation behind this move is that HadCM3 has been observed to model the climate very well, giving realistic approximations to surface temperature, and meridional heat transport (Pope et al., 2000). Additionally, spatial patterns of the quantities shown in equations 1.3 and 1.4 all exhibit the features of the climate that are consistent with present day observations. Owing to its realistic representation of the climate, HadCM3 has been a popular choice for model researchers.

We note that the \( \cos \) latitudinal weighting on surface area is not included for the spatial patterns, however it is taken into account for the zonal mean distributions, such as fig. 1.3. The sign convention is such that positive (negative) implies into (out of) the system.

Figure 1.1: Schematic of incoming solar radiation incident at the Earth’s surface. For identical intensities of radiation per unit area through A and B, intensity at the surface on A will be almost the original value, whilst the intensity at B will be much lower since the same energy is spread over a larger area. Figure is taken from Neelin (2011).

Incoming SW solar radiation (Fig. 1.2A) from the sun is the main forcing on our climate system. Due to the spherical geometry of our planet it follows a symmetric distribution about the equator, decreasing towards the poles (see Fig. 1.1). Due to the Earth’s surface not
being a perfect absorber, a part of this SW radiation is reflected back into space as outgoing SW radiation (Fig. 1.2B). This reflected radiation happens at all locations on the planet however it is particularly strong within Antarctica, Greenland and high orographic features (mountains) where the presence of snow and ice on these areas reflects a higher portion of radiation. Snow and ice surfaces can have an albedo of up to $\approx 0.8 - 0.9$ (Allison et al., 1993). Other high reflectance regions include much of the mid-latitude belts ($30^\circ - 60^\circ$) where there are high concentrations of cloud cover that act as natural reflectors. Regions of low reflectance include the Tropical belts ($5^\circ - 30^\circ$) where the presence of cloud cover is relatively low. The open ocean surface has excellent absorption properties with an albedo of $\alpha \approx 0.1 - 0.3$ (Jin et al., 2004). Outgoing LW radiation (Fig. 1.2C) has a maximum in the equatorial and tropical regions where the largest surface temperatures are located, with much less at higher latitudes. Due to all these effects the net TOA radiation (Fig. 1.2D) is such that there is a net surplus of absorbed radiation within the Tropics, and a net deficit at higher latitudes. The net distribution thus implies an advection of heat from the Tropics towards the Poles to maintain an equilibrium.

Figure 1.2: Spatial distribution of radiation at the top of the atmosphere for A). Incoming SW radiation, B). Outgoing SW radiation, C). Outgoing LW radiation, and D). net radiation. Data is taken from the HadCM3 control run, years 80-119. Units are in $W m^{-2}$ without cos latitudinal area weighting.

Figure 1.3 illustrates the zonal mean distribution of the above radiation quantities. Comparing magnitudes, there is a fine imbalance between the large fluxes of longwave and shortwave radiation, with the net accounting for a very small amount. Aside, this is still enough energy
to drive both the ocean and atmospheric circulations as we observe.

At the ocean-atmosphere interface, the presence of these two media adds two extra components to the energy balance at the surface. The first is termed the *Sensible* heat flux ($S$), and is proportional to the temperature difference between the surface and the air directly above it. If cold air is travelling above a warm surface, there will be a cooling effect on the surface and a loss of thermal energy, with the reverse being true. The sensible heat flux can be written as $S = c_o M \Delta T/A \delta t$, where $c_o$ is the specific heat capacity of seawater, $M$ is the mass of seawater (in GCMs this would be the top layer of ocean), $\Delta T$ is the temperature difference between an increment of time $\delta t$, and $A$ is the surface area of the region. The second is termed *Latent* heat flux ($L$) and describes the energy loss due to surface evaporation. In this mechanism the energy lost by the ocean is proportional to the energy required for a given flux of freshwater to remain within a vapour state. Latent heat flux can be calculated from $L = l_v F_v/A$, where $l_v$ is the enthalpy of vaporization of water, and $F_v$ vertical freshwater flux out of the region. Both $S$ and $L$ vertical heat fluxes are in units of $Wm^{-2}$.

We can thus describe the surface energy balance equation by

$$R_{surf} = SW_{net} - LW_{net} - S - L \tag{1.4}$$

where the ‘net’ subscripts define the net radiation for the SW and LW components. As previously, the spatial distribution of the quantities in Eq. 1.4 is shown in Fig. 1.4, with the corresponding zonal mean latitudinal distributions in Fig. 1.5. A landmask is added to only highlight the ocean surface; although land surfaces are important to climate physics, it non-advective properties and high surface albedo makes the land component an almost negligible
medium for heat transport, warranting its neglect. In equilibrium, $R_{surf} \approx 0$ over land.

Much of the oceanic net absorption of SW radiation occurs within the Tropical Latitudes (Fig. 1.4A) leading to the greatest warming. As a result the largest net emittance of LW radiation happens within the same region (Fig. 1.4B).

Within the top 1000m of the ocean, large-scale advection mechanisms are predominantly in the form of gyres, driven by surface wind stress (see section 1.4.4). For the Northern Hemisphere Tropics, these clock-wise lateral circulations transport equatorial warm waters to higher latitudes. Much of this northward transport of warm water happens in an intense jet of water known as Western Boundary Currents, due to their location of the western side of the ocean basin. As such, the water transported at these western locations have much higher temperature than the zonal average, leading to an outward flux of sensible heat (Fig. 1.4C). The other large sensible heat flux signals found near Greenland are not reflective of gyre behaviour but are due to the intense westerly winds that blow over these regions from continental North America. Particularly during winter time, these extremely cold winds can lead to intense outgoing sensible heat fluxes aiding in deep water formation (Pickart et al., 2003). Vertical Latent heat flux (Fig. 1.4D) follows similar distributions to sensible heat flux and outgoing LW radiation, being found in warm regions where surface evaporation can occur. Note that latent heat flux is practically zero polewards of 60° since the temperatures are cold and evaporation is low. The net sea surface radiation (Fig. 1.4E) thus describes an intense absorption of radiation within the equatorial region, and localised outlets of energy situated near the major western boundary and deep water formation regions (all situated in the northern hemisphere), and to a smaller extent, the Southern Ocean Region, where warm ocean eddies formed within the Aghulas Current (off the coast of South Africa) are shed and advected into the region. From this representation we would therefore expect a symmetric poleward ocean heat transport but with a much larger northern hemisphere component than for the south (see Fig. 2).

### 1.3 Atmospheric Circulation

The Earth’s atmosphere is a gaseous mixture of mainly Nitrogen (78%) , oxygen (21%) and a few other trace gases that is evenly distributed across the planet. Though the atmosphere can reach a height exceeding 100km, more than 99% of its mass is found below 30km (Peixoto and Oort, 1992). Due to gravity, the atmosphere is stratified in layers, with the densest (lightest) layers the bottom (top). The high compressibility and low density make the atmosphere a
Figure 1.4: Spatial distribution of radiation at the top of the atmosphere for A). Incoming SW radiation, B). Outgoing SW radiation, C). Outgoing LW radiation, and D). net radiation. Data is taken from the HadCM3 control run, years 80-119. Units are in $W m^{-2}$ without cos latitudinal area weighting.

Figure 1.5: Zonal mean distribution of quantities listed in Fig. 1.4 with cos(latitude) weighting.

highly dynamical and unstable fluid.
1.3.1 Dynamics of the Atmosphere

All motion can be described by Newton’s 2nd law of motion,

$$\vec{F} = m\vec{a} \quad (1.5)$$

where $\vec{F}$ is the force on a mass $m$, and $a$ is the acceleration on that body.

By dividing Eq. 1.5 by mass and decomposing acceleration, we achieve

$$\frac{d}{dt} \text{velocity} = F_{\text{cor}} + F_{\text{pgf}} + F_{\text{grav}} + F_{\text{drag}} \quad (1.6)$$

where we have used the time derivative of velocity as acceleration; forces on the right hand side are per unit mass to have the same units as acceleration. The pressure gradient force $F_{\text{pgf}}$ and $F_{\text{drag}}$ relate to frictional forces such as the turbulent drag between parcels, or the frictional drag near a surface boundary. $F_{\text{grav}}$ represents the gravitational force that acts to pull mass towards the Earth. The Coriolis force, $F_{\text{cor}}$, is due to the rotation of the Earth and appears from differences between reference frames. A travelling parcel observed on a rotating plane will seem to travel a different distance when observed from a stationary plane. The Coriolis force is defined as

$$\vec{F}_{\text{cor}} = -2\overline{\vec{\omega}} \times \vec{u} \quad (1.7)$$

where $\vec{F}_{\text{cor}}$ is the Coriolis force per unit mass, $\overline{\vec{\omega}}$ is Earth’s angular velocity, and $\vec{u}$ is the velocity vector of the object in the rotating frame ie. Earth’s rotating frame. The Coriolis Force acts in the perpendicular direction to travel, deflecting objects towards the right (left) in the NH (SH).

We define the Coriolis parameter $f$ such that

$$f = 2\pi\omega\sin(\theta) \quad (1.8)$$

where $\theta$ is latitude. Since the value of $f$ is dependent on latitude, it is sometimes easier to define the latitudinal change of $f$, known as the $\beta$ parameter,

$$\beta = \frac{1}{R} \frac{df}{d\theta} \quad (1.9)$$

where $R$ is the radius of the Earth. $\beta$ is proportional to the $\cos(\theta)$ and is always positive and maximum at the equator.

The Earth receives more solar radiation per unit area at the equator than near the pole. This causes a non-uniform heating of the planet with the equator being warmer (higher pressure) than the poles (low pressure).
Consider two different pressures at positions $x_1$ and $x_2$ respectively. If the pressure at $x_1$ is greater than at $x_2$ then there will be an acceleration towards the lower pressure region. If both pressures are the same, then there is no net acceleration in any direction. By taking a small increment in the $x$ direction between positions $\delta x$, there will be corresponding incremental change in pressure $\delta p$. The pressure gradient force along the $x$ axis is therefore $\delta p/\delta x$, with units of force per unit volume. To gain units of acceleration, we divide by density $\rho$ such that the complete pressure gradient force per unit mass is $-\frac{1}{\rho} \frac{\delta p}{\delta x}$, with a similar equation in the $y$ direction.

As such, there is an ascendance of warm, moist air at the equator which travels poleward at height. As the air ascends it cools and precipitates its water vapour, thus the equatorial region has a net precipitation to evaporation rate. To balance out the top branch of this circulation there must be a lower circulation of cold air from the poles towards the equator to complete the loop. For a non-rotational body this convective cell would stretch from the equator towards the poles, however due to the Earth’s rotation this type of circulation would be impossible to maintain. We shall see below how the balance of forces, particularly the Coriolis force, shapes the atmosphere’s circulation structure.

By use of the above equations, the horizontal velocity equations can be written as

$$\frac{du}{dt} = fv - \frac{1}{\rho} \frac{\partial p}{\partial x} + F_{\text{drag}}^x$$  \hspace{1cm} (1.10a)

$$\frac{dv}{dt} = -fu - \frac{1}{\rho} \frac{\partial p}{\partial y} + F_{\text{drag}}^y$$  \hspace{1cm} (1.10b)

where $u$ and $v$ correspond to east-west and north-south horizontal velocities, $p$ is pressure, and $F_{\text{drag}}^x$ and $F_{\text{drag}}^y$ correspond to turbulent drag of the flow in the $x$ and $y$ direction respectively. This term arises from small scale turbulent processes that usually cannot be resolved in climate models, and thus have to be parameterised. They also include the effects of surface drag that tend to slow wind speed down and accelerate ocean currents.

The time mean meridional overturning circulation of the atmosphere is shown in Fig. 1.6c with each hemisphere having three overturning circulations - these are the tropical Hadley Cell, the midlatitude Ferrel Cell, and the Polar cell; the idealised single circulation cell schematic for a non-rotating body thus vanishes when rotation is taken into account. In the Hadley Cell there is an ascendance of warm, light air at the equator with a subsidence (sinking) of cold, dense air at the sub-tropics which leads to a thermally driven direct circulation. In the Ferrel Cell there is an ascendance of cold air at high latitudes and a descendance of warm air at mid-latitudes leading to a thermally indirect circulation (Peixoto and Oort., 1992).
latitudes are dominated by large-scale transient disturbances that transport a vast amount of heat polewards; the Ferrel Cells are only the residual circulation, schematised by the time mean motion. The Polar Cell is a weak, direct cell.

Figure 1.6: Zonal mean cross sections of a). the zonal wind speed component in \( \text{ms}^{-1} \), b). the meridional wind component in \( \text{ms}^{-1} \), and c). the annual and zonal mean atmospheric mass streamfunction in \( 10^{10} \text{kgs}^{-1} \).

Taken from Oort and Peixoto (1984).

Just polewards of the Hadley Cells there exist extremely fast (25 \( \text{ms}^{-1} \)) zonal wind speeds that occur at high altitudes. By using simple mathematics we explain the mechanics of the Hadley Circulation, and why there are such high zonal wind speeds at mid-latitudes.

**Thermally Driven Circulations**

In a steady column of fluid, the dominant balance of forces on a section of fluid is between gravity pushing it downwards, and the pressure gradient force pushing it back up such that

\[
\frac{\partial p}{\partial z} = -\rho g
\]  

(1.11)

This is known as hydrostatic balance. By combining Eq. 1.11 with the pressure gradient force \( \frac{1}{\rho} \frac{\partial p}{\partial z} \), the latter is replaced by \( g(\partial z/\partial x) \), the same being analogous in the \( y \) direction. This simple equation has far reaching consequences when simplifying climate physics, the most important being the use of pressure as a vertical coordinate. Since \( gz \) is the gravitational potential energy per unit mass, we can define this as the geopotential such that

\[
\Phi = gz
\]  

(1.12)
The *Equations of State* (EoS) links the density of a substance to mainly its temperature, pressure and other such factors. In the atmosphere, the EoS is known as the *Ideal Gas Law*, defined by

\[ \rho = \frac{p}{RT} \]  

(1.13)

where \( R = 287 \text{Jkg}^{-1}\text{K}^{-1} \) is the gas constant for dry air, and temperature \( T \) is in Kelvin. Density is proportional to pressure, with higher pressure resulting in more molecules per unit volume. It is inversely proportional to temperature with higher temperatures leading to expansion therefore decreasing molecule density per unit volume.

By combining the Ideal Gas law with the Hydrostatic equation by substituting Eq. 1.13 in Eq. 1.11, gives

\[ \frac{\partial \ln p}{\partial z} = -\frac{g}{RT} \]  

(1.14)

where \( \frac{\partial \ln p}{\partial p} = 1/p \). Integrating Eq. 1.14 in the vertical direction and assuming that temperature is roughly constant in the Kelvin scale, gives

\[ p = p_0 e^{-z/H} \]  

(1.15)

where we define \( H = \frac{RT}{g} \) as the scale height. The former relationship explains why pressure decreases with height, albeit it, at a non-linear rate. The negative exponential fall in pressure shows us that pressure changes become increasingly smaller at high altitudes, making it very difficult to define the ‘top of the atmosphere’.

If we instead multiply Eq. 1.14 by \( RT \) and integrate in the vertical, by using 1.12, we simply get

\[ \Phi - \Phi_0 = \int_p^{p_0} RT d(ln p) \]  

(1.16)

where \( \Phi_0 \) is the geopotential height at pressure \( p_0 \). Eq. 1.16 explains that the difference in height between two geopotential surfaces is proportional to the temperature between the two, since warmer air leads to an expansion of the air column and therefore an increase in height. The above equations can be applied to explain the mechanisms of thermally driven circulations in the atmosphere that are particularly prevalent in the Tropics.

Fig. 1.7 shows an idealised setup for a thermally driven circulation. We consider two geopotential heights \( z_1 \) and \( z_2 \) (black dashed lines) that stretch over a latitudinal region where
Figure 1.7: Schematic of a thermally driven circulation cell. Over a warmer surface, the air directly above will be thicker compared to over a cold region, causing ascendance of fluid (dashed red arrow). The difference in column thickness bends constant density isopleths (solid red lines). Relative to planes of constant geopotential height (dashed blue lines), an area of high pressure at \( z_2 \) forms over the warm region, with a similar low pressure over the cold region at the same height. This causes a pressure gradient force that advects fluid from A to B at height. At heigh \( z_1 \), the pressure gradient is reversed, with fluid being advected from B to A nearer to the surface. Figure taken from Neelin (2011).

there is a warm surface on one end (A), and a colder surface on the other (B). In the real world this can represent the Tropics and Extra-tropics respectively. If surface temperature was uniform, the two surfaces of constant atmospheric pressure i.e. isobars (solid red lines) would be parallel to surfaces of constant \( \Phi \). Above the warm surface region, heating of air leads to an expansion of surrounding air, resulting in lighter parcels relative to its surroundings that start to convect upwards. The warm air also leads to an overall expansion in the atmospheric column, pushing the isobars at \( z_1 \) and \( z_2 \) further apart. Over the cold region (relative to the warm one), the air is denser, leading to a contraction of the air column, which simultaneously pulls the isobars closer together. At \( z_2 \), there is thus a region of high pressure for the warm region relative to the cold one, driving a pressure gradient force polewards from A to B. At the lower altitude of \( z_1 \) the reverse is true, with high pressure over the cold region and low pressure over the warm region, driving an equatorward transport or air, from B to A. The strength of the circulation is proportional not only to the temperature gradient between the two regions, but also the thickness of the regions (between isobars) at which the temperature differences are occurring.

This is a highly idealised schematic, but is essentially the mechanisms for driving the north-south Hadley Cells over the Tropics, and the east-west Walker Circulation over the Pacific that vary according to the zonal oscillations of warm surface waters near the equator that are related to El Niño/ La Niña events.
Thermal Wind Equation

The ascendance of moist air at the equator forms part of the Hadley Cells poleward moving flank. Due to the Coriolis force, air parcels are deflected to the right (left) in the NH (SH). At its poleward limit of around 30° air parcels are advected in a near perfect zonal direction at altitude. The balance of forces are predominantly between the Coriolis Force and the pressure gradient force. In absence of surface wind drag, the velocity equations of Eq. 1.10a and Eq. 1.10b can be rewritten as

\[ f u_g = -\frac{1}{\rho} \frac{\partial p}{\partial y} \]  \hspace{1cm} (1.17a)

\[ f v_g = \frac{1}{\rho} \frac{\partial p}{\partial x} \]  \hspace{1cm} (1.17b)

where the subscript \( g \) represents the Geostrophic Velocity. This also explains why regions of low pressure spin anti-clockwise, whilst high pressure regions spin clockwise in the northern hemisphere. Realistically, geostrophic balance is never perfect and does not work near the equator where the Coriolis Parameter converges to zero.

As represented in Fig. 1.6a wind speeds can reach up to 25 \( ms^{-1} \). These are obviously faster than the wind speeds nearer to the ground, however apart from the effect of surface drag, why would wind speeds be so large at such high altitudes? To probe this question we combine the Hydrostatic Equation (1.11) with the Geostrophic equation (1.17a and (1.17b).

\[ f \frac{\partial u}{\partial z} = -\frac{g}{T} \frac{\partial T}{\partial y} \]  \hspace{1cm} (1.18a)

\[ f \frac{\partial v}{\partial z} = \frac{g}{T} \frac{\partial T}{\partial x} \]  \hspace{1cm} (1.18b)

These are the Thermal Wind equations and describe how the increase in velocity with height is proportional to the perpendicular gradient of surface temperature below. Since the mid-latitudes are regions containing the highest meridional temperature gradients, the area above must have extremely fast wind speeds aloft according to the Thermal Wind equations.

Again, following the Hadley Cell’s equatorward moving flank, the Coriolis Force again deflects motion to the west, creating the equatorial Easterly (from the east) Winds, that are also known as the Trade Winds. The convergence of equatorward motion from both hemispheres feeds the upward motion of moist air, completing the loop.
1.3.2 Longitudinal Distribution of the Atmosphere

The latitudinal distribution of the Atmosphere’s circulation shown in Fig. 1.6c was zonally averaged. In reality, atmospheric processes in the east-west direction can vary as much compared with in the north-south direction. The main reason for this is the presence of different land configurations, allowing for varying surface heating rates that in turn affect the climate above. Continental structures change the flow of ocean currents that would otherwise be zonally symmetric on an aqua planet configuration.

![Maps of the annual mean distribution of evaporation E (top), precipitation P (middle) and E-P (bottom). Green areas represent net evaporation whilst brown areas represent net precipitation. Figure taken from Marshall and Plumb (2008).](image)

Figure 1.8: Maps of the annual mean distribution of evaporation E (top), precipitation P (middle) and E-P (bottom). Green areas represent net evaporation whilst brown areas represent net precipitation. Figure taken from Marshall and Plumb (2008).

Fig. 1.8b illustrates the time mean precipitation rates from NCEP re-analysis data over a 40 year period (Kalnay et al., 1996). Since the ocean is itself, a large source of freshwater to the atmosphere, the largest precipitation rates occur over Open Ocean, with smaller amounts located over land. Precipitation rates are particularly low within the tropics where the descedance of dry air leads to excess evaporation that result in desert surroundings over land, and extremely salty water over the ocean.

The inter-tropical convergence zones (ITCZ) are characterised by an intense rain belt...
that circumvents the entire globe. It’s annual mean position lies slightly northward from the equator however there is a seasonal oscillation where the ITCZ moves northward (southward) in northern (southern) summer, particular over land. These oscillations go hand in hand with the Monsoon Circulations that have been traditionally linked with local changes in wind direction.

Over the major ocean basins, precipitation rates are larger over the western side rather than the eastern side. Western boundary moisture enhancement couples strongly to the locations of the very warm western boundary currents found in the ocean. These current regions release a large amount of latent heat (moisture) from the surface, with precipitation rates usually being substantially larger over them than the surroundings. In the time mean, the western side of an ocean basin is attributed to a large amount of vertical moist convection, with the eastern side associated with descending motion. Similar to the north-south Hadley Cells, this describes a zonal circulation known as the Walker Circulation, shown in Fig. 1.9.

![Figure 1.9: Schematic of the zonal atmospheric circulation patterns found within the Tropics. Deep convection regions over south-east asia and descending motion near Panama relate to the Walker Circulation that occurs over the Pacific. Figure taken from Marshall and Plumb (2008) which has been adapted from Madden and Julian (1972), and Webster (1983).](image)

1.4 Structure of the Ocean

1.4.1 Temperature Distribution

The sea surface temperature (SST) of the ocean is shown in Fig. 1.10. The largest SSTs are found in the tropical belt with values decreasing poleward, following the spatial distribution of solar heating. The largest meridional temperature gradients are found at mid-latitudes, particularly in the Southern Ocean ($40^\circ S - 70^\circ S$) where there are no land masses to inhibit water flow. For these reasons, SSTs are often very nearly zonally symmetric, except at coastal boundaries due to the presence of land mass and atmospheric influences. In subtropical basins, SSTs are usually warmer at the western side than the eastern side, due to the presence of
western boundary currents such as the Gulf Stream (Kuroshio) in the North Atlantic (North Pacific) bringing in warm water from the deep tropics poleward (see Wind Driven Circulation). On the eastern side of the North Atlantic sub-polar gyre, SST are also warmer than the zonal average due to the north-easterly direction of the Gulf Stream bringing in its warm water towards the British Isles and continental Europe. The anomalously cold SSTs on the eastern side of the equatorial basins are due to the coastal upwelling of cold, nutrient rich waters from the deep ocean abyss, fuelled by the easterly Trade winds in the region.

![Figure 1.10: Annual mean values of sea surface temperature in °Cs. Warm (cold) waters are represented by darker (lighter) shades. Figure taken from Marshall and Plumb, (2008), which uses data from the Levitus World Ocean Atlas (1994).](image)

In the vertical plane, the warmest waters are found at the surface, while the coldest waters are found in the deep ocean. Without any atmospheric winds, isotherms (planes of constant potential temperature) would smoothly outcrop at the ocean’s surface the further poleward, creating a ‘u’ shaped, due to differential surface heating. The shape of the isotherms is similar to that of the letter ‘w’, due to the influence of the atmospheric winds (see Fig. 1.11). Upwelling in the equatorial and sub-polar gyre regions due to Ekman vertical motion bring up water from below and therefore bend the isotherms towards the surface, whilst downwelling in the sub-tropical gyre regions depress the isotherms, sinking them further downwards. The bending of these isotherms (more to do with the bending of isopycnals, however temperature and density are well correlated within the upper layer) are a major source of potential energy for the ocean as the bringing up of dense water and the depression of light water is an unstable configuration that requires energy to maintain. This potential energy is normally dissipated by oceanic eddies that try and flatten out the isotherms (Gent and McWilliams, 1998; Jayne and Marotzke, 2002) thus returning the ocean to a more stable energy configuration.
Below the surface of the ocean the temperature structure can be divided into 3 main layers: a relatively well mixed layer at the surface, a region with rapid temperature decrease with depth, and a stable, cold region (see Fig. 1.12). Wind induced turbulence at the surface of the ocean induce mixing of water properties, such as temperature and salinity to a depth of between 50-200m (Pickard and Emery, 1990), depending on season and latitude. Below this depth there is a region of rapid temperature decrease (largest temperature gradient) called the Thermocline where the diffusion of heat from above is halted. Due to seasonal variations in heating and the errors in oceanographic measurements, it is easier to determine a thermocline range for which it should exist rather than its exact depth. In the Tropics and mid-latitudes, the Thermocline depth is between 200-1000m and is referred to as main Thermocline. At higher latitudes the transference of heat from the ocean into the atmosphere breaks the stratification of the ocean in which column water density can become well mixed all the way towards the deep interior. Since these events happen seasonally and due to the outcropping of isopycnals.
near the surface, there is no main Thermocline at these latitudes. Below roughly 1000m the temperature profile remains almost constant with depth at 2-3 $^\circ$C with very little seasonal variation in temperature - this is the deep abyssal layer.

**Figure 1.12:** Idealised ocean potential temperature profile with depth. The surface mixed layer has a fairly homogenous temperature, owing to turbulent mixing due to the atmosphere. Below there is a transition where temperature decreases rapidly with depth, known as the Thermocline. Below this layer is the Abyssal layer which has a very homogenous temperature of a few degrees above freezing point. Figure is taken from Marshall and Plumb (2008).

### 1.4.2 Salinity Distribution

The Salinity of the ocean refers to its salt concentration. It is calculated as the ratio of the mass of salt in a fixed volume (in grams) of sea water to the total volume mass (in kg). The unit of salinity is the *practical salinity unit*, or psu, and has dimensions of g/kg. As an example, a salinity of 34 psu would correspond to 34 g of salt diluted in 1 kg of water. Actual measurements of salinity are much more complicated, relying on the electrical conductivity of water to derive salinity values.

Sea surface salinity (SSS) has a very near zonal distribution just like its temperature counterpart (Fig. 1.13). SSS is tightly coupled with evaporation E, and precipitation P; the former increasing surface salinity while the latter decreases it. The surface salinity minimum can be found just north of the equator where there is net P over E i.e. E-P is less than 0, corresponding to the precipitative rich, ITCZ region. SSS maxima are found at 25° where E-P is greater than 0 i.e. a net evaporation over precipitation. Anomalous low salinity values can also be found near coastal regions where net freshwater run-off from land and rivers help
dilute surface salinity. Other places include Polar Regions, particularly around melting ice which also dilutes water from freshwater melt.

![Figure 1.13: Annual mean values of sea surface salinity in psu. Salty (fresh) waters are represented by darker (lighter) shades. Figure taken from Marshall and Plumb (2008), which uses data from the Levitus World Ocean Atlas (1994).](image)

The vertical profile of salinity is much more complex than the stratified temperature distribution. Waters with different density re-arrange to the lowest energy configuration possible, such that the densest waters are below and the lightest waters are on top (see Fig. 1.14). Density is both a function of temperature and salinity. At high SSTs, temperature fluctuations override salinity fluctuations in determining density. Therefore it is possible to see both low and high salinity waters at the surface.

Between the equatorial and tropical regions there is a marked salinity minimum between 600m to 1000m, with an increase to stable values at 2000m. In the tropics there is usually a sharp increase in salinity between 100m - 200m which comes from the sinking of dense water at the SSS tropical salinity maximum which eventually flows equatorward. At coastal areas where there is much river-runoff there is usually a region where the salinity increases rapidly with depth, normally termed the *halocline* (Pickard and Emery, 1990).

Within the abyssal depths of below 2000m salinity values are relatively uniform at 34.6 to 34.9 psu making these waters very uniform throughout the whole ocean.

### 1.4.3 Density Distribution

The density of seawater $\rho$ is defined as the mass per unit volume. It is a complex function, reliant on both non-linear quantities of temperature, salinity, and pressure. Incidentally, this
Figure 1.14: Annual mean values of salinity in psu for the world’s oceans. The top panel illustrates the 1 km layer from the surface, whilst the bottom shows the total depth. Saltier (fresher) waters are represented by darker (lighter) shades. Figure taken from Marshall and Plumb, 2008, which uses data from the Levitus World Ocean Atlas (1994).

is the EoS of the ocean, defined by

$$\rho = \rho(T, S, P)$$

(1.19)

Density is measured in $kgm^{-3}$ and range from 1002.00 $kgm^{-3}$ on the open ocean surface to about 1029.00 $kgm^{-3}$ in the deep ocean (Marshall and Plumb, 2008) By convention, it is the density anomaly that is stated, calculated as the density difference from pure water $\rho_0 = 1000$ $kgm^{-3}$. Since seawater is extremely hard to compress, the pressure dependency on density is usually ignored, being set to a prescribed value. Given all these corrections, we define the density anomaly term $\sigma_t$ that is a function of only temperature and salinity,

$$\sigma_t = \rho - \rho_0$$

(1.20)
Figure 1.15: Isopycnals surfaces of constant density as a function of both potential temperature and salinity. The dashed box represents the T-S space in which 90% of ocean waters resides in, and includes also the global mean T and S. The freezing point is represented by the dashed line while the temperature of maximum density is the straight solid line. Figure adapted from Pickard and Emery (1990).

To understand ocean density better, its distribution in temperature and salinity space is represented in Fig. 1.15.

The solid curve lines are surfaces of constant density, also known as isopycnals. The solid straight line is the temperature of maximum density, whilst the dashed straight line is the freezing point. 90% of ocean water resides in striped T-S box (which also resides the mean T and S point for the world ocean).

As can be seen from the graph, $\sigma_t$ varies significantly at high temperatures for all values of salinity, however these variations decrease the lower the temperature, particularly at low values of salinity. The changes in $\sigma_t$ with salinity remain fairly constant throughout the whole range for T and S. The temperature of maximum density of freshwater is about 4 °C and decreases linearly with increasing salinity until about 26 psu where maximum density and freezing point have the same temperature. The freezing point of water also decreases linearly with increasing salinity, reaching a minimum of -2 °C. These coldest waters are formed off the coast of Antarctica and the polar North Atlantic, where brine rejection due to ice formation and cold water temperatures lead to some of the saltiest waters (and usually the densest) waters in the world that sink into the deep abyss to form Antarctic Bottom Water (AABW).

Since salinity is a conserved quantity in the ocean it can be used as a tracer for individual water masses to map out the deep circulation of the ocean. For the Atlantic Ocean there are 4 major waters namely the North Atlantic Deep Water (NADW), Antarctic Bottom
Water (AABW), Antarctic Intermediate Water (AAIW) and North Atlantic Common Water (NACW), shown in Fig. 1.16. These waters form part of the vast internal structure of the ocean and quantifying how the location of these waters change with climate change could be important in understanding future climate states.

Figure 1.16: Cross sectional profile of density through the Altantic Basin. Atlantic water masses can be traced through the isopycnal contours: NACW floats above the \(35 \, kgm^{-3}\) contour in the northern hemisphere, below at the \(34.8\) and \(34.9 \, kgm^{-3}\) lies the AADW. In the southern hemisphere the tongue of water starting at Antarctica and sinking until 1000m is AAIW, while the \(34.7 \, kgm^{-3}\) contour falling into the deep abyss is AABW. Figure adapted from Marshall and Plumb (2008)

1.4.4 Wind Driven Circulation

The wind driven circulation is primarily observed in the first few hundred metres of the ocean surface. On large scales frictional forcing on the surface can induce gyral motions in the horizontal plane with length scales of a few 1000km, whilst in the vertical, motion is communicated downward through successive stratified layers.

Ekman Transport

The atmospheric circulation imparts some of its motion on the ocean through surface windstress. Through shear forces between ocean layers, this surface momentum is transferred downwards into the ocean. The friction between each layers acts as a dampener on vertical momentum transmission such that below a certain depth, the transport of vertical momentum is zero. This depth range over which surface windstress effects the ocean is termed the *Ekman Layer* (Ekman, 1905). The net mass transport in the Ekman layer can be derived from the horizontal velocity equations of Eq. 1.10a and 1.10b. By assuming a constant flow where
variations to \( u \) and \( v \) in time are minimal, we can rewrite these as

\[
-fu + \frac{1}{\rho_0} \frac{\partial p}{\partial x} = F_{x}^{\text{drag}} \tag{1.21a}
\]

\[
fu + \frac{1}{\rho_0} \frac{\partial p}{\partial y} = F_{y}^{\text{drag}} \tag{1.21b}
\]

where \( \rho_0 \) is a reference density set at 1000 \( kgm^{-3} \) and all other terms have their usual meaning. If we consider the drag terms as force exerted per unit mass then we can approximate the latter by

\[
F_{x}^{\text{drag}} = \frac{1}{\rho_0} \frac{\partial \tau_x}{\partial z} \tag{1.22a}
\]

\[
F_{y}^{\text{drag}} = \frac{1}{\rho_0} \frac{\partial \tau_y}{\partial z} \tag{1.22b}
\]

where \( \tau_x \) and \( \tau_y \) are the horizontal wind stress components of \( \vec{\tau} \). The Ekman component (or wind-driven) of the horizontal momentum equation can be written as

\[
\hat{k} \times \vec{u}_{ek} = \frac{1}{\rho_0 f} \frac{\partial \vec{\tau}}{\partial z} \tag{1.23}
\]

where \( \hat{k} \) is the unit normal vector. By integrating the latter equation over the Ekman layer depth \( h_{ek} \) and multiplying by \( \rho_0 \), yields the Ekman layer mass transport,

\[
\tilde{M}_{ek} = \int_{h_{ek}}^{0} \rho_0 \vec{u}_{ek} dz = \frac{1}{f} \hat{k} \times \vec{\tau}_{surf} \tag{1.24}
\]

where \( \vec{\tau}_{surf} \) is the surface wind stress. Components pertaining to the meridional and zonal mass transport can be written as

\[
M_{x}^{ek} = \frac{1}{f} \tau_y \tag{1.25a}
\]

\[
M_{y}^{ek} = -\frac{1}{f} \tau_x \tag{1.25b}
\]

In the NH Ekman transport is always towards the right of the wind stress vector, whilst it is to the left in the SH. A consequence of horizontal advection within the Ekman layer results in vertical motion. By the continuity equation,

\[
\frac{\partial u_{ek}}{\partial x} + \frac{\partial v_{ek}}{\partial y} + \frac{\partial w_{ek}}{\partial z} = 0 \tag{1.26}
\]

where \( w_{ek} \) is the vertical velocity within the Ekman layer. By assuming that the vertical velocity at the surface is zero, we can calculate the vertical velocity at the Ekman layer base \((-h_{ek})\) by vertically integrating \( w_{ek} \) over the layer depth, leading to

\[
w_{ek}(-h_{ek}) = \frac{1}{f \rho_{ref}} \text{curl}(\vec{\tau}_{surf}) \tag{1.27}
\]
where we represent \( \text{curl}(\vec{\tau}) = \hat{k} \cdot \nabla \times \vec{\tau} \). If the wind stress curl is cyclonic (i.e. anti-clockwise) then Ekman dynamics dictate a vertical upwelling of water. Similarly if motion is anti-cyclonic (clockwise), vertical motion is then downwards. These two motions are more commonly referred to as **Ekman Suction** and **Ekman Pumping** respectively and are very important especially for ocean gyres. To illustrate this process, fig. 1.17 gives an idealised schematic of the zonal mean wind stress. The sub-tropical gyre is flanked by westerly winds above and easterly winds below, resulting in anti-cyclonic motion, whereas the sub-polar gyre spins in the opposite direction. The differences in upward motion tend to bend isopycnal slopes of the ocean interior that adds available potential energy to the ocean. Atmospheric winds thus communicate motion and energy down into the ocean through surface interactions.

The Sverdrup Relation

The Sverdrup Relation is a very important concept in fluid dynamics that arises due to the rotation of the Earth. Vorticity is a measure of spin about a vertical axis. **Planetary Vorticity** describes the Coriolis Parameter \( f = 2\Omega \sin(\phi) \) and is equal to double the Earth’s angular momentum multiplied by the sin of latitude. **Relative Vorticity** \( \xi \) is defined by the curl of horizontal velocity, expressed by

\[
\xi = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \tag{1.28}
\]

**Absolute Vorticity** \( \eta = f + \xi \) is the summation of planetary and relative vorticity, with lastly **Potential Vorticity** defined as \( \eta \) divided by the height of the water column \( H \). Put simply, Sverdrup Balance describes the conservation of potential vorticity. As an example, if relative...
vorticity increased, conservation of potential vorticity would require the fluid column to either
move southwards, or stretch.

By taking the curl of the horizontal momentum equations (1.10a and 1.10b) and using the
continuity equation 1.26, we obtain

$$\beta v = f \frac{\partial w}{\partial z}$$  (1.29)

which is known as Sverdrup Balance.

**Western Intensification**

The pattern of zonal wind stress patterns on the ocean’s surface results in basin wide gyral
circulation such as the sub-tropical and sub-polar gyres in the Atlantic and Pacific regions.
Observations of ocean currents show an intensification of northward moving waters near the
western side of ocean basins with a broader but slower return flow on the eastern side. Why
the strongest currents are only observed on the western side is explained below.

We modify the horizontal velocity equations by writing them as

$$fv = \frac{1}{\rho_0} \frac{\partial p}{\partial x}$$  (1.30a)

$$fu = -\frac{1}{\rho_0} \frac{\partial p}{\partial y} - rv$$  (1.30b)

where we have assumed geostrophic balance. The extra term $-rv$ represents a frictional
force term of the basin coast which increases proportional to velocity. There is no $-ru$ term
in Eq. 1.30a since water flowing east-west have no coastal areas for friction to occur. By using
the product rule of differentiation on the above and eliminating the pressure terms with the
continuity equation, the Sverdrup balance equation is also modified into

$$\beta v + r \frac{\partial v}{\partial x} = f \frac{\partial w}{\partial z}$$  (1.31)

At the surface we assume no vertical motion; the RHS of eq. 1.31 equals to zero, ie.
$$\beta v + r \frac{\partial v}{\partial x} = 0.$$ The solution to this equality becomes $v \propto e^{-\frac{\beta v}{r} x}$. Northward velocity is
therefore greatest at the western side and decreases with eastward distance from the coast.
Eastern intensification therefore cannot exist since any solution moving from the eastern side
into the interior would *increase* in velocity - the opposite of what is observed.
1.4.5 Thermohaline Circulation

The terms *thermo* and *haline* refer to temperature and salinity characteristics of the ocean. From the previous section on density characteristics, salinity variations only become important in the cold temperature regime. The Thermohaline Circulation (THC) therefore refers to the deep abyss of the ocean with its origins starting at high polar latitudes. Counter to its name, the THC has both a buoyancy driven side at high northern latitudes and a wind-driven component in the Southern Hemisphere. This duality often gets referred to as the ‘push’ and ‘pull’ mechanisms of the THC (Radko et al., 2008).

The THC is a slow and sluggish process of overturning in the ocean and is started by vertical sinking of dense water, sinking at high latitudes and spreading horizontally in the deep interior. Although evaporation in the tropics does lead to an increase in salinity, the increase in temperature due to solar heating overrides this mechanism and no sinking of water takes place at low latitudes. Sinking of water therefore takes place at high latitudes through two known mechanisms - near boundary sinking and Open Ocean sinking.

Near boundary sinking primarily happens in the Weddel and Ross seas off the coast of Antarctica (Fig. 1.19, orange circles) Brine rejection from ice formation increases the salinity of the water on top of the Antarctic continental shelf. Eventually the accumulation of heavy, salty water slides off the continental shelf where it mixes with cold, upwelling AADW to form the densest waters of the Atlantic, called AABW (see Fig. 1.16).

Open sinking is a phenomenon that happens at the Labrador Sea in the North Atlantic, and off the coast of eastern Greenland. At these locations sinking is done through large surface cooling rates. These waters accumulate and spill over the sills between Greenland and Scotland and enter the North Atlantic.

**Stommel and Arons Model**

One of the earliest theories to characterise the deep abyssal circulation was proposed by Stommel (1958) and Stommel and Arons (1960). These works predicted an equatorward divergence of deep water away from their high latitude source regions, below the wind-driven western boundary currents. To close the loop of these Deep Western Boundary Currents (DWBC), Stommel argued that the poleward return flow would occur within an upper layer, brought about by a uniform upwelling of water near the equator. This circulation is shown in Fig. 1.18. By using Sverdrup Balance, the flow of water due to upwelling can therefore be pre-
dicted. Integrating Eq. 1.29 in the vertical we retrieve

$$\int_D^z v dz = \frac{f}{\beta} w(z)$$  \hspace{1cm} (1.32)

where $D$ is the depth of the ocean basin, and $z$ is an arbitrary mid-depth level. The model assumes that $w(D) = 0$ i.e. no vertical velocity at the ocean floor. Since $w(z)$ is always positive, water columns will stretch. Due to conservation of vorticity these will move polewards towards the source regions.

There have been observations of DWBC in the real ocean (Bower and Hunt, 2000a,b) however the symmetric picture of deep abyssal flow according to Stommel and Arons model has been shown to be inconsistent with observations (see Doos and Coward, 1997). Interior flowing North Atlantic Deep water has been observed to flow southward, extending down towards the Southern Ocean where upwelling takes place (Fig. 1.19).

**Figure 1.18:** Schematic of the abyssal ocean circulation according to Stommel. Source regions of deep water are represented by black dots within the North Atlantic and Southern Ocean, flow equatorward as Deep Western Boundary Currents. Return flow is achieved at a higher level through upwelling near the equator. Figure taken from Stommel (1958).

### 1.4.6 Conceptual Models of the Thermohaline Circulation

One of the earliest and simplest models used to understand the THC was proposed by Stommel (1961) and is shown in Fig. 1.20a. The model consisted of two well mixed boxes of identical volume, each with unique temperature and salinity, representative of low (box 2) and high (box 1) latitudes. Both boxes were connected to each other by flow paths near the surface and at the base, with interchange between the two boxes possible at a flow rate of $q$. This was derived by the density difference between the two boxes given by

$$q = k[\rho_1 - \rho_2] = k[\alpha (T_2 - T_1) - \beta (S_2 - S_2)] = k[\alpha \Delta T - \beta \Delta S]$$  \hspace{1cm} (1.33)
Figure 1.19: Schematic representation of the Thermohaline Circulation: surface currents are shown in red, deep waters in light blue, and bottom water in dark blue. The major convection sites are the orange circles and highlight the Labrador Sea and The Greenland Sea in the North Atlantic, and the Ross and Weddel Seas off the Antarctic Coast. Figure adapted from Rahmstorf (2002).

For this model setup it was found that there were two stable models of equilibrium. The first was when temperature variations dominated any changes to density, resulting in \( q > 0 \) and a poleward near surface flow. The second mode was when salinity variations dominated changes to density, leading to an opposing flow that aimed to reduce \( q \). Put simply, the ‘salinity’ mode can be seen as a braking mechanism for the THC whilst the ‘temperature’ mode enhances it. Changes to salinity for both boxes were affected by freshwater exchanges at the surface whereby net evaporation (precipitation) led to an enhancement (freshening) of salinity. In the model this was represented by an equivalent surface salinity flux \( H \).

Figure 1.20: a. Stommel’s Two box model from 1961, b. Booth’s 3 box model from 1982. All boxes are assumed to be well mixed with unique values of temperature and salinity. Coloured arrows represent flow between boxes whilst black arrows signify salinity fluxes \( H \) at the surface. Flow strength represented by the variable \( q \). Figures taken from Marotzke (2000).

A more sophisticated version of Stommel’s box model was developed by Rooth (1982) which now took into account both hemispheres (see Fig. 1.20b). Rooth’s model consisted of high
latitude regions for the NH and SH (box 1 and box 3 respectively), and a tropical region (box 2) shared by both hemispheres. Flow is now assumed to be controlled by a pole-to-pole density gradient, with equivalent surface salinity fluxes in both hemispheres and a symmetric tropical temperature gradient across the equator. It was found that the THC was only dependent on the SH atmospheric moisture flux (Rahmstorf, 1996), and that its circulation strength was proportional the freshwater forcing. This second point is in direct contrast to Stommel’s two box model where an enhanced hydrological cycle led to a reduction in THC strength. Finally, the model predicted that the model’s stable state of $q > 0$ would become unstable if $H_N/H_S > 4$ (Scott et al., 1999). The latter meant that if freshwater input at high northern latitudes became too much, then it would lead to a THC collapse. These conceptual models and their implications have fuelled much research into the possibility that (sudden) changes to the THC may have affected past climate (and vice versa), such as the paleo-climatic Younger Dryas (Manabe and Stouffer, 1988). Presently, the focus has shifted to understanding how the THC will change when forced with anthropogenic changes (Solomon et al., 2007).

### 1.5 Climate and Climate Change

Over the last few decades, the human race has begun to gradually understand that our way of life is having a profound influence on the climate in which we live in. As a result, scientific communities and government agencies have been working together to understand the mechanisms that affect our climate. Future predictions to idealised greenhouse gas scenarios still have considerable spread. Warming surface temperatures will have a profound effect on both ocean and atmospheric circulations that will change the patterns of freshwater transport. In this section we describe and discuss the principal mechanism underpinning climate change, in the form of the Greenhouse Effect. We shall then proceed to identify three main indicators of climate change, namely: surface temperature change, atmospheric weather patterns changes, and ocean circulation changes. Prior to this however, a brief overview of the Earth’s historic climate over the last 4.5 billion years is discussed as a perspective to compare to today’s present climate.

#### 1.5.1 Earth’s Climate History

The Earth’s climate has never been constant, but has in the past changed dramatically, swinging between extreme cold events such as the periodic Ice Ages, to the ice free paleoclimatic period of the Eocene. The existence of these past climate states have led to a new branch of
study, known as *paleo-climatology* whose main aim is to study and understand Earth’s past climatic history. Such studies are building up an increasingly detailed picture to different types of mechanisms, feedbacks and tipping points involved, particularly to abrupt climate change events. Presently there is an urgent need to know how our way of life will impact our climate in the near future. The last studies from the Intergovernmental Panel on Climate Change (Solomon et al., 2007), predicted increases in global mean surface temperatures due to anthropogenic forcing, however the large spread in model errors is still a major issue. The analysis of past climate states thus acts as a great test bed to understand mechanisms associated with e.g. abrupt climate change, that will help as indicators to our present time.

Our earliest records of the Earth’s climate spans approximately 4.56 Ga (Baker et al., 2005), the earliest originating from proxy examination of sedimentary strata. The chemical composition of these ancient rock fragments could tell us the early composition of the early Earth’s atmosphere.

From the climate record we can generally separate different regions into two sets of climate states: *greenhouse* state and *icehouse* state.

Certainly around the early Archaen period (see Fig. 1.21), oceans were present and the temperature held to a narrow range for the presence of liquid water. Earliest temperatures at this period range from 50-80 °C before cooling to 20-30 °C (Robert and Chaussidon, 2006). Much of the precambrian period (4500-550 Ma) equated to warming phase with the atmosphere being made up of high concentrations of carbon dioxide and methane, both potent greenhouse gases that have been suggested as the main means of maintaining such as high temperature (Ohmoto et al, 2004). There were however striking icehouse events puncturing the period whereby Earth’s temperature dropped dramatically. Recently these have been linked to the popular *Snowball Earth* hypothesis (Budyko, 1969; Kirschvink, 1992) whereby the entire globe was covered with ice, terminating any ocean-atmospheric interactions completely (Hoffman et al., 1998). A theoretical mechanism for the Snowball Earth hypothesis was given by Budyko (1969) which stated that if ice-cover extended further equatorward than 30 °, then the ice-albedo feedback would take over, extending ice cover completely to the poles. However the idea of a completely covered world has been widely challenged, with a preference for a partially covered world, or *Slushball* Earth (Fairchild and Kennedy, 2007).

Some of these extreme events have been linked to the natural astronomical variations in the orbital characteristics of the Earth, for instance its eccentricity (how stretched the orbital path is), axial tilt, and precession (orbital ‘wobble’), resulting in very low frequencies of approx-
Figure 1.21: Historical record of Earth’s surface temperature at a range of different time-scales. ‘T’ denotes relative temperature. Figure taken from Letcher (2009)

imately 100, 30 and 20 ka respectively (Zachos et al., 2001). These led to periodic changes in amount of solar radiation absorbed by the planet. When amplified by various feedback mechanisms, in particular greenhouse gas concentrations, the results were the observed glacial patterns in the past (Milankovitch, 1941).

The Holocene period started roughly 15 ka ago (Fig. 1.21 panel f) and is associated with a period of very stable temperatures and sea-level especially when compared to other previous interglacial periods. Climate variations have been estimated to be on the order of 1 °C and include such examples as the ‘Medieval Warm Period’ and ‘Little Ice Age’.

Following the man’s modern industrial revolution of about 200 years ago, there has been a steady increase in fossil fuel use that has led to increases in greenhouse gases, particularly $CO_2$ concentrations (Keeling, 1960). Indeed this has led to the concept of the *Anthropocene*, or geological period dominated by human induced activity (Crutzen, 2002), and the resulting termination of the Holocene (Zalasiewicz et al., 2008).
1.5.2 Understanding Global Warming

The ability to warm the atmosphere simply by increasing certain gases that were sensitive to absorbing radiation is not a recent discovery, but in fact a concept that has been known for almost 400 years. An overview of the history of the Greenhouse effect is given in the IPCC AR4, Chapter 1.

One of the earliest ideas for the Greenhouse Effect was observed by Edme Mariotte in 1681, noting that although the sun’s radiation easily passed through objects such as glass, heat from other sources could not. In the 1760s an experiment by Horace Benedict de Saussere used a pane of glass covering a thermometer in a darkened box, providing at least a conceptual idea to the Greenhouse Effect. This idea was extended by Fourier (1824) and Pouillit (1836), whom made the link that air could absorb thermal radiation. In 1861 John Tyndall researched the thermal absorption properties for a number of gases found in the atmosphere. He found that the largest atmospheric gases (by volume) eg. Nitrogen and Oxygen, were not very absorbing, unlike trace gases such as $CO_2$, $O_3$ and $H_2O$ which were. Tyndall noted that all past climates could have been reproduced simply by changing the atmospheric concentrations of these high thermally absorbing Greenhouse Gases (GHG).

The author thus defines the Greenhouse Effect, described in the context of our climate as, the artificial warming of the surface and lower troposphere through an increase in concentration of highly absorbing and re-radiating gases of thermal radiation.

Callendar (1938) solved a set of equations that linked GHG to changes in the climate. By his calculations he found a doubling of $CO_2$ concentrations would lead to an increase in global mean surface temperatures by approximately 2 °C with a faster rate of heating near the poles. Callendar’s work is both surprising and impressive since he documents changes that are in agreement to the predictions of the most sophisticated climate models to date (see Chapter 2 for details).

Revelle and Suess (1957) studied the carbon cycle with relation to the interactions of the ocean and atmosphere with $CO_2$ concentration. They found that although the upper ocean could rapidly absorb atmospheric $CO_2$, the rate at which this $CO_2$ is passed into the deep ocean is on the order of a few centuries. Although the ocean does play an important part in moderating $CO_2$ concentrations, the majority thus stays in the atmosphere.

By the 1970s, a wider variety of GHGs other than $CO_2$ and $H_2O$ were discovered, namely, $CH_4$, $N_2O$, and CFCs (Ramanathan, 1975; Wang et al., 1976. In the same decade the effects of aerosol cloud effects also became known (Twomey, 1977).
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The concentration of GHGs over the last century have risen at a steady rate (Keeling, 1960). If one is to believe the above scientific observation on GHGs, then a warming of the Earth’s climate is already under way. In the next sections we describe three main indicators to global warming.

1.5.3 Indicators - Sea Temperature

The global oceans cover roughly two thirds of the planet’s surface with an average depth of 4.3 km\(^1\). Such a huge mass of water is an essential mediator of Earth’s climate given that seawater has a heat capacity three orders of magnitude larger than that of the atmosphere (Bindoff et al., 2007). Over the last 40 years, 84% of the climate’s warming has gone into the ocean (Barnett et al., 2005). Over the previous 50 years, the ocean’s energy content has increased by approximately \(14.2 \times 10^{22}\) J, compared with less than \(1 \times 10^{22}\) for the atmosphere (Bindoff et al., 2007). These figures suggest a build-up of heat in the ocean that would otherwise accumulate in the atmosphere. The ocean thus acts as a giant heat reservoir that aids in damping out large increases in Earth’s global mean temperature.

Traditionally there have been two main methods of measuring sea temperature. The first is sea surface temperature, measured across the first few metres of the ocean, whilst the second is the heat content, integrated across a large depth of approximately 3000m (Letcher, 2009, Chapter 19). Historically, sea temperature was measured by passing ships travelling the ocean. The problem of this method was the lack of both spatial and temporal resolution for any meaningful global mean estimates to be obtained. With advent of drifting buoys in the 1970s and the satellite era of the 1980s, global coverage increased significantly.

![Global mean surface temperature anomalies](http://oceanservice.noaa.gov/facts/oceandepth.html)

Figure 1.22: Global mean surface temperature anomalies based on analysis by Thompson et al. (2009). Vertical lines represent the time period of August 1945 and major volcanic eruption dates. From left to right these represent Santa Maria, Mount Agung, El Chichón, and Mount Pinatubo. The horizontal line represents the time mean surface temperature for the period 1961-1990.

\(^1\)taken from http://oceanservice.noaa.gov/facts/oceandepth.html
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Figure 1.22 shows the raw SST data from the International Comprehensive Ocean-Atmosphere Data Set (ICOADS, Worley et al., 2005) based on work by Thompson et al. (2008). There was observed a cooling phase at the early state of the 20th century before rising to a peak during the 1940s. Apart from a smaller cooling period after World War 2 (WW2), the temperature trend increases throughout the last half of the century up until present day. The top figure displays corrected values of SST corrected for the use of un-insulated buckets prior to the 1940s (Thompson et al., 2008). This has the added effect of slightly raising the SST records before 1941, however still in line with an overall increase in temperature. The large discontinuity near 1945 has been attributed to a change in measurement methodology that came into effect after WWW2 (Letcher, 2009, Chapter 19).

The cooling phase during the early part of the 20th century has been linked to the volcanic eruptions of Krakatoa (1883) and Santa Maria (1902) which injected a vast amount of aerosols into the atmosphere that acted as a cooling mechanism for the entire planet (Forrest and Reynolds, 2008). Indeed, volcanic eruptions have accounted for many large dips in temperature, the latest being Mt. Pinatubo (1991, Church et al., 2005); see Fig. 1.23.

Comparison of the upper-ocean heat content (black; grey shading indicates an estimate of one standard deviation error) with previous estimates (red and blue) for the upper 700 m. The straight lines are linear fits to the estimates. The global mean stratospheric optical depth (beige, arbitrary scale) at the bottom indicates the timing of major volcanic eruptions. The brown curve is a three-year running average of these values, included for comparison with the smoothed observations. In the lower figure, comparison of the 700-m (thick black line, as in a) and 100-m (thick red line; thin red lines indicate estimates of one standard deviation error) results with sea surface temperature (blue; right-hand scale). All time-series were smoothed with a three-year running average and are relative to 1961.

The last half century has seen a large variety of variability in the ocean SST record, particularly at the decadal scale, owing the effects of the ENSO cycle (Letcher, 2009, Chapter 19). The El Niño event of 1997-1998 (McPhaden, 1999) was the largest of its kind on record, leading to an increase of 0.2°C above the average temperature (Hansen, 2006) and culminating in the spike in SST records for the same period. The return of the cooler La Niña period in the early 2000s has possibly resulted in the levelling off of the temperature record for the last decade.

Although there has been a steady increase in global mean surface temperatures (Knutson et al., 2006), there have been assymetric differences with the Northern Hemisphere increasing
in temperature \((0.71 \pm 0.06^\circ\text{C})\) much faster than for the Southern Hemisphere \((0.64 \pm 0.07^\circ\text{C},\) Rayner et al., 2006). Even more so on a regional scale, the Arctic has been warming faster than the global average since 1966 (Lindsay and Zhang, 2005). On the Eastern side of the tropical Pacific, the warming trend is somewhat modest compared to the entire basin (Rayner et al., 2006), due to the increased Trade Winds that lead to regional coastal upwelling (Cane et al., 1997). This has led to an enhancement of the zonal temperature gradient across the equatorial Pacific that is generally seen as a precursor to El Niño events (McPhaden, 1999). A gradual warming has been observed in all major basins that was out of range of temperatures just forced by natural variability. (Barnett et al., 2005, their Fig. 2). Studies have shown that the Atlantic warms progressively throughout the first 700m of the ocean (Barnett et al., 2005). For the Pacific and Indian oceans, warming is confined to the top 100 m, with little or no changes at further depths, and even cooling near 150 m. The different characteristics have been attributed to the deep convection sites in the Atlantic that allow for a much deeper
mixing. Whereas for the Pacific and Indian ocean, deep convection does not occur (Warren, 1983; Emile-Geay et al., 2003) with heating committed only to the shallow surface circulations.

In summary, there has been evidence of considerable warming of the oceans since 1940, particularly with marked increases since the 1970s. Anthropogenically induced surface warming has been observed throughout all major basins though there is considerable internal variability due to the different basin circulations. The higher increases in Arctic warming from Polar Amplification across the entire Northern Hemisphere will lead to considerable feedbacks into the atmosphere, most notably a slowdown in its circulation.

1.5.4 Indicators - Atmospheric Circulation

In this section we describe the main changes of the atmospheric circulation based on available observational data and model studies. An outline of the atmosphere’s general circulation patterns is provided in Section 1.3.

We focus primarily on the poleward expansion of the tropical region with a corresponding poleward shift to the sub-tropical jets.

Poleward Expansion of the Tropical Circulation

In the past there have been many indicators for the location of the separation point between the Tropics and the Extra-Tropics. These can be split up into two groups of dynamical indicators, and physical indicators. The first describes characteristic features of the atmospheric circulation such as the poleward edge of the Hadley Cell or the position of the sub-tropical jets. The second describes certain aspects of the atmosphere with sharp lateral gradients e.g. outgoing LW, and evaporation minus precipitation.

One of the earliest studies that focused on the width of the Tropics was presented by Rosenlof (2002) who studied the latitudinal extent of the Brewer Dobson Circulation in the lower stratosphere. This describes a slow meridional circulation that extends between the troposphere and stratosphere, with upwelling in the Tropics and downwelling further polewards. By using this indicator Rosenlof calculated an increase of 3 ° per decade for the period 1992-2001. Later, Reichler and Held (2005) used the tropopause height as an indicator, with the sharp latitudinal gradient of the high altitude Tropics and low altitude extra-tropics. Using a combination of reanalyses and atmospheric radiosonde data they found that the Tropics were expanding by approximately 0.4 ° per decade since 1979, substantially smaller than Rosenlof. The same study focused on the separation distance between the two sub-tropical jets, becom-
ing wider with time. A similar study by Seidel and Randel (2007) based on the tropospheric height as an indicator yielded similar results of tropical expansion. Hudson (2006) looked at the distribution of ozone between the tropics and extra-tropics. From the use of spectrometers they found that the low ozone concentration region in the Northern Hemisphere, characteristic of the tropics had increased over time.

Estimates of tropical widening range between 0.3-3.0 °, with an average of about 1.4° (Reichler, 2008). The wide range of estimates reflects the observational uncertainties and methodologies used. Some estimate, e.g. 3° widening per decade may perhaps be too large since these would produce shifts in the climate system that have not already been observed.

Evidence from models has been able to reproduce the widening of the tropics seen in observations when forced with 20th century climate scenarios. For the ensemble models of the IPCC AR4 report, there was shown a widening of the tropopause height (Reichler and Held, 2005). The largest model showed an increase of 0.7° per decade over the last 30 years, however other models produced substantially smaller expansion rates, and some even produced contractions. Other robust consequences of the tropical expansion have been documented; Kushner et al. (2001) observed a poleward shift of the sub-tropical jets using a GCM forced with increasing CO₂ concentrations at 1% per year. The intensive CO₂ scenarios (A2) from the IPCC AR4 report also produced similar results with an ensemble mean shift of approximately 0.2° per decade scenarios (Lorenz and DeWeaver, 2007).

Model based calculations do seem to support present observations, however the rate of tropical poleward expansions is seen as too smaller, with even the severest of climate scenarios. Some have argued that the under-representation by models may be due to the model’s poor representation of atmospheric features, such as the stratosphere (Seidel et al., 2008).

**Decrease in Intensity of the Tropical Circulation**

There has been theoretical arguments (Knutson and Manabe, 1995; Held and Soden, 2006) that the tropical circulation will slow down in a global warming environment. A key response of the atmosphere to increases in CO₂ is that water vapour content increases at a rate such that relative humidity RH is held constant (Wetherald and Manabe, 1975; Held and Soden, 2006; Lorenz and DeWeaver, 2007). Evidence to support this notion has been verified from observation data at interannual timescales (Trenberth et al., 2005; Soden et al., 2005). The reason’s as to why RH is taken as a constant is not straightforward (see Pierrehumbert et al., 2007) given that most of the atmosphere is far from saturated. In the free troposphere, atmo-
spheric circulation patterns act to maintain a relatively dry atmosphere through precipitation from upward convection. The RH of an air parcel in the free troposphere can be understood through tracing backwards in time to the point of last saturation (Held and Soden, 2000; Pierrehumbert et al., 2007). Provided that we neglect the mixing between air parcels and considering their trajectories, it was shown that the assumption of fixed RH was equivalent to saying that the change in temperature of an air parcel from last saturation is similar to the change in the air parcel itself (Held and Soden, 2000). The assumption of fixed relative humidity therefore happens when trajectories are constant and that temperature changes are uniform. Given that the latter two conditions are not met in the real world, the fact that relative humidity is approximately constant would imply that these changes are of secondary importance when dealing with water vapour content in the atmosphere (Lorenz and DeWeaver, 2007).

The specific humidity of air $q$ is defined as the mass ratio of water vapour to dry air in a specific volume. The saturation specific humidity $q_s$ is the maximum amount of water vapour that a specific volume can hold before condensation occurs. This quantity is dependent on temperature and can be expressed by

$$\frac{dq_s}{dT} = \frac{l_v}{R_v T^2} q_s$$  \hspace{1cm} (1.34)$$

where $l_v$ is the latent heat of vaporisation, $R_v$ is the Gas Constant for water vapour at constant volume, and $T$ is temperature. By integrating (1.34) between temperatures $T_2$ and...
and taking the exponential we get

\[
\frac{q_s(T_2)}{q_s(T_1)} = e^{\exp \left[ -\frac{l_v}{R_v} \left( \frac{1}{T_2} - \frac{1}{T_1} \right) \right]}
\]

(1.35)

Multiplying by \(q_s(T_1)\) eq. 1.35 can be expressed as

\[
q_s(T_2) = q_s(T_1) \exp [\alpha (T_2 - T_1)]
\]

(1.36)

where \(\alpha = l_v/R_v T_1^2\). By taking relative humidity \(RH = q/q_s\) as a constant, we can finally express eq. 1.36 as

\[
q_2 = q_1 \exp (\alpha \Delta T)
\]

(1.37)

where we have dropped the temperature dependence notation. The above equation thus states as a consequence of fixed relative humidity, changes to specific humidity with increasing temperature will change at an exponential rate, known as the Clausius-Clapeyron, or C-C rate (Boer, 1993; Allen and Ingram, 2002; Held and Soden, 2006). \(\alpha\) has a value of approximately 0.07, simply meaning that specific humidity increases by about 7% for every 1°C increase in temperature.

The increase in specific humidity is the result of increased evaporation and precipitation rates, and an intensification of the hydrological cycle. If all else was held constant i.e. the atmospheric circulation, then the transport of moisture would also increase (Lorenz and DeWeaver, 2007). It so happens that due to energetic constraints of the climate, the increase in the hydrological cycle is less than the C-C rate. The reason to why this was the case was researched by Allen and Ingram (2002). They concluded that the overall intensity of the hydrological cycle was not controlled by the amount of moisture in the atmosphere, but by the ability of the troposphere to radiate away latent heat released through precipitation. In a simplified schematic we look at the radiation balance at the surface and the tropopause. Net incident radiation at the surface is a summation of incoming incident solar radiation and a smaller upward cooling component back into space. At the tropopause, net radiative cooling \(\Delta R\) (directed into space and towards the surface) is balanced by an upward flux of latent heat release \(l_vP\) (\(l_v\) is enthalpy of vaporisation of water, and \(P\) is global mean precipitation), thus evaporation cools the surface whilst precipitation heats the troposphere.

Using the same methodology of Allen and Ingram (2002), we model the radiative perturbation at the tropopause as

\[
\Delta R_C + \Delta R_T = l_v \Delta P
\]

(1.38)
where we have decomposed the change in net radiation $\Delta R$ belonging to a component, $\Delta R_C$, that is independent of processes associated with the global mean surface temperature change $\Delta T$, and processes that are dependent, $\Delta R_T$. The radiative perturbations associated with $l_v \Delta P$ have been found to be about 1 Wm$^{-2}$ for a 1% increase in global mean precipitation (Allen and Ingram, 2002). As explained above, changes to $\Delta R_C$ are independent of variations to $\Delta T$. To understand this quantity better and serving as an example, a doubling of atmospheric CO$_2$ would result in a decrease in terrestrial radiation out into space by 3-4 Wm$^{-2}$, however increase downward radiation to the surface by about 1 Wm$^{-2}$ (Mitchell et al., 1987), resulting in a net value of $\Delta R_C$ between -2 to -3 Wm$^{-2}$. Thus if there is no changes to the temperature gradient (ie. $\Delta R_T = 0$) we would expect an overall decrease in the hydrological cycle ie $\Delta P < 0$. This was shown in early modelling studies using prescribed SST values (Mitchell, 1983) that were held constant.

The increase in radiative cooling from a warming troposphere, $\Delta R_T$ is approximately proportional to $\Delta T$. That is, changes to tropospheric temperature scale roughly to variations at the surface such that $\Delta R_T = k_T \Delta T$, with $k_T \approx 3$ Wm$^{-2}$K$^{-1}$ (Mitchell et al., 1987). It is thus this balance between $\Delta R_C$ and $\Delta R_T$ that constrains the increase of the hydrological cycle to a rate less than C-C. A consequence of this was highlighted by Held and Soden (2006) whom stated that as a result, one would expect the atmospheric circulation to reduce with increasing CO$_2$. Held and Soden examined this issue by approximating atmospheric mass transport as $M \propto p/q$, where P is global mean precipitation, and q is the global mean specific humidity. If q increases at the C-C rate and precipitation at a lesser rate, the atmospheric circulation must therefore slow down to compensate for this difference. Results from the latest IPCC report do show model evidence for an atmospheric slow-down, at least for the tropical Hadley Cell (Held and Soden, 2006). The decrease over the tropical region has been attributed to the reduced meridional temperature gradient, especially for the northern hemisphere (Ma et al., 2011 [submitted to J. Climate]).

Real world observations of the zonal Walker Circulation over the Pacific from sea level pressure measurements show a decreasing trend in circulation strength (Vecchi et al., 2006; Zhang and Song, 2006). Model simulations under global warming scenarios have shown similar trends (Tanaka et al., 2004; Vecchi and Soden, 2007), with the likely mechanism being the anomalous warming of SSTs (Tokinaga et al., 2011).

Models suggest that the weakening of the tropical circulation affects the zonal Walker Circulation much more than the meridional Hadley Cell (Lu et al., 2007). Studies done from
atmospheric reanalysis data (Mitas and Climent, 2005, 2006) show a somewhat mixed picture with little or no slowdown, and also an intensification in the circulation.

1.5.5 Indicators - Ocean Circulation

We have described how the ocean circulation can be decomposed into the wind-driven and THC components (section 1.4.4). Although the former component contains the fastest moving waters, it only comprises the upper 1000 m of the ocean. By its shear spatial structure the THC is an obvious choice to examine when analysing the overall circulation of the ocean, in particular the AMOC circulation of the Atlantic region.

The AMOC is the largest meridional overturning circulation cell in the entire ocean, responsible for transporting a large amount of heat northwards, even within the Southern Hemisphere. The ‘pushing’ of the circulation at high northern latitudes due to the deep convection provides ventilation of the deep ocean, whilst the ‘pull’ in the Southern Hemisphere from westerly wind-stress brings up deep water towards the surface.

As discussed in section 1.4.5 the strength of the THC can be seen as driven by density gradients, reflecting variations in temperature and salinity. The variations in the strength of the AMOC have been shown to correlate well with deep water formation in key convection sites (Dong and Sutton, 2005). The formation rate of these dense waters has the potential to slow down provided that there are favourable conditions; an increase in surface temperature leads to lighter waters at the top of the water column with colder waters successively below, enhancing stratification. If freshwater input at the surface is increased, whether from enhanced precipitation or from continental runoff and ice melt, this again acts to decrease surface density. Both effects therefore act to subdue downward vertical convection.

During the last Ice Age (100 000 - 10 000BC), Greenland experienced periods of rapid warming of about 10 °C within a few decades, followed by a gradual cooling of a few millennia (Johnsen et al., 1992). These so called ‘Dansgaard-Oeschger’ events (Dansgaard et al., 1993) have been linked to sudden changes in the AMOC circulation that in turn, affected the surrounding climate (Adkins et al., 1997). Warm phases of the North Atlantic are thought to be the result of a strong AMOC, with cold phases being linked to a weak AMOC state, described by the near collapse of NADW formation. During the last Ice-Age, the subsequent weakening of the AMOC has been assumed to have been caused by a sudden freshening event of the high North Atlantic region by the Laurentide Ice Sheet (Hemming, 2004). This would have led to stronger stratification of the water column leading to the inhibition of deep convection.
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and the formation of NADW. Numerical evidence to support such an event exists. Through
the use the HadCM3 model, Vellinga and Wu (2004) caused a shutdown of the AMOC by
artificially adding a freshwater flux at high latitudes. This ‘Hosing’ Experiment resulted in an
overall temperature decrease of about 8 °C over the Greenland area, with relatively modest
decreases for the rest of the Northern Hemisphere. Paleo-climate records have also suggested
that changes to the AMOC during the last deglacial period (19000 - 14500 years ago) were
mirrored by a significant input of ocean CO₂ into the atmosphere, raising concentrations by
50 ppm (Sarnthein et al., 2007).

In the IPCC AR4 report (Solomon et al., 2007) it concludes that there is a ‘very likely’
chance that the AMOC will slow down by the end of the 21st century by about 0-50% based
on the greenhouse gas emission scenario A1B. These are shown in Fig 1.25.

![Figure 1.25: Predicted trends of the strength of the Atlantic Meridional Overturning Circulation at 30 °N
from an ensemble of coupled climate models forced by the SRES A1B emissions scenario. Figure taken from
IPCC AR4, Chapter 10.](image)

The large spread in model predictions amount to the unique complexities of each model,
whether that be differing model resolution, grid type, vertical coordinate system, parameteri-
sations but to name a few.

The physical mechanisms of the AMOC slowdown included the high latitude input of
freshwater from melting Arctic and Greenland Ice, the intensification of the hydrological cycle
(Pierrehumbert, 2002), and the decrease in net surface heat flux cooling, that all act to stabilise
the water column from vertical motion. From our mention of Stommel’s Two-Box model in
section 1.4.5, input of high latitude freshwater could enhance the ‘haline’ component of the
THC, leading to a weakened circulation. In conjunction we would expect a reduced ‘thermal’
component of the THC from polar amplification, with the temperature difference between low
and high latitudes decreasing with CO₂ concentration.
Though we list a few of the main mechanisms, the model spread in predictions to the slowing of the AMOC in Fig. 1.25 still leaves much to be desired from GCM models.

Other studies have found stabilising responses of the AMOC to increasing global temperatures. Model studies by Latif (2000) and Thorpe et al. (2001) found that anomalous high surface salinity waters would develop in the Atlantic Tropics from an increased freshwater transport from the Atlantic to the Pacific. These would be advected northwards, aiding in deep water convection, partially stabilising the AMOC slowdown due to freshening. The ‘slowing-down’ effect originating at high latitude has a complex regional picture, with different mechanisms affecting both the Labrador Sea and the Greenland-Iceland-Norway region (Hu et al., 2004).

On the ‘pull’ side at the Southern Ocean Huang et al. (2006) found an increase in wind stress across the region between 1950-2000 leading to an increased northward Ekman transport. The possible cause has been hinted at the decreasing stratospheric ozone concentration (Thompson and Solomon, 2002). Model observations by Shindell and Schmidt (2004) was able to reproduce the increased wind stress over the Southern Ocean region, forced by anthropogenic warming.

Direct observations of the strength of the AMOC have been heavily reliant on hydrographic sections of density to implicitly measure mass transport from geostrophy. From five such measurements at 24.5 °N over a fifty year period, showed an AMOC reduction of 8 Sv, or roughly 30% (Bryden et al., 2005). Studies of dense water inflows through the Faroe Bank Channel near the Nordic seas have seen a decrease of 1-2 Sv since 1970 (Hansen et al., 2001).

A new hydrographic section has been proposed at 26.5 °N whereby continuous monitoring of the AMOC strength would take place (Cunningham et al., 2007; Kanzow et al., 2007. The variability of the AMOC over a 1 year period measured a time mean of 18.5 ± 5.6 Sv (one standard deviation limit) with a range of about 30 Sv between maximum and minimum values (Cunningham et al., 2007). The large annual variability of the circulation would suggest a difficulty in statistically proving the observed reduction of 30% over a 50 year period. The need of a longer data set is therefore required for any trends to be detected (Baehr et al., 2007).

1.6 Summary

The Earth’s climate is a highly complex system governed by the motions of the ocean and atmosphere, and driven by the Sun’s solar radiation. For millennia, it has been this energy
that has sustained the planet’s temperature to ranges hospitable for life to flourish. It has only been very recently that we have discovered that human induced behaviour has been contributing to a secondary warming of the climate with profound feedbacks for the ocean and atmospheric general circulation. The most noticeable changes have been the increase in surface temperatures, the poleward migration of the Tropics and the sub-tropical jets, and a (possible) slowing down of the AMOC (according to the latest ensemble model predictions), but to name a few.

The next chapter deals with numerical models used in this report. We will discuss the theoretical considerations that constrain many models, along with their strengths and weaknesses, with particular relevance to future climate predictions.
In this Chapter we describe the basis of climate models, concentrating first on the main components (equations) and how they are computationally solved. We next describe three main models used for research purposes. These are the global general circulation models (GCM) \textit{HadCM3} and \textit{CHIME}, and a simple two-box model named \textit{EPcm}. A description of the different climate scenarios used and important definitions are given after, along with a brief summary.

2.1 The Construction of a Climate Model

GCMs have become an essential tool for both simulating our climate, whether it be synoptic weather patterns that can be predicted for up to a few days, to decadal and centennial predictions on human induced climate change. Although many models have similarities in what they are calculating, all are fundamentally different in complexity, model structure, resolution, parameterizations etc. Due to processing power available, there is usually a compromise between the resolution of a model and the time available for ‘running’ the model; the latter, alternatively described as model spin-up, or model integration.

Continuous fields of climate variables i.e. temperature, specific humidity, velocity etc. are represented by a finite amount of values. Computationally this is called as \textit{discretisation} where a discrete value of a property is chosen as the average over a particular region. This brings in the concept of model \textit{resolution} whereby the higher the number of discrete values for a given
variable (e.g. globally) leads to a better spatial representation. There exist many methods of discretising continuous data, which will be discussed further into the section.

This section discusses the basics of how simple coupled climate models are built, the key parameterisations and their limits.

2.1.1 An Atmospheric Model

![Figure 2.1](image)

Figure 2.1: a) An atmospheric model complete with topographical features and grid representation b). an individual atmospheric layer taken from the above picture c). schematic of the individual fluxes into and out of a single grid cell box. Figure taken from Neelin (2011).

A typical representation of an atmospheric model is shown in Fig. 2.1 complete with topographical representation of mountains. A single layer is then taken for simplicity.

Pressure is used as the vertical coordinate with increments $\Delta p$. Typical values of $\Delta p$ can range between 10-150 hPa depending on vertical height above the surface (see next paragraph for details). Similarly, the horizontal axis constitute longitude and latitude with corresponding increments $\Delta x$ and $\Delta y$ respectively. Typical ranges for the horizontal resolution of GCMs can be between $2 - 5^\circ$, calculated to approximately 200-500km. For regional climate models this resolution is typically higher (up to $0.1^\circ$) since the spatial area of interest is much smaller than for an entire global coverage. The shaded regions represent a single grid cell; circles denote grid boundaries whilst arrows represent calculated fluxes into and out of the box. For each
grid box there exists only one value for each variable. Any processes that require a resolution smaller than that of the grid box or comparable to, are not fully resolved and must therefore be parameterised.

To aid in modelling efficiency, the vertical pressure coordinate resolution is not constant, increasing for tropospheric layers and being much coarser for higher altitudes. Horizontal resolution may change however the vast majority of models do not, using a constant grid spaced representation. The Mercator map method of equally spaced vertical meridian lines converging at the poles, results in singularities and vanishing surface areas. To get over this problem modellers have developed hybrid model grids that locate the poles over land masses (see later for details). In this configuration both horizontal resolution can vary with location. This method is particularly useful for ocean models where processes near the true geographical poles need to be efficiently modelled e.g. Arctic Ice/Ocean dynamics.

The partial differential equations of motion are replaced by equations that involve the differences in variables between adjacent grid cells e.g. horizontal wind velocity is computed by the vertical pressure difference between adjacent cells consistent with geostrophy. Each box communicates with its nearest neighbour through the advection of quantities i.e. energy, wind, moisture, salinity etc. Fluxes are proportional to the differences of each variable between adjacent grid cells. From considering the fluxes of properties into and out of a grid cell, a rate of change can be computed for each variable. This is used to compute a new value of e.g. pressure than in turn computes a new value of velocity after one time step. This sequence is reiterated for all grid boxes so that the new values are available for the next time step. Typically the temporal resolution is about half an hour or less (Neelin, 2011). A model is spun-up until the necessary duration being modelled is achieved.

For any model, the treatments of sub-grid processes must be taken into account and not solely rely on the explicit calculations of the large scale. This is particularly true for the atmosphere where local processes occurring on spatial scales < 100km are out of range of its horizontal resolution. As an example, we explain the treatment of moist convection within the atmosphere. This process is non-local, meaning that individual grid points do not just interact with adjacent points, but all points within the atmospheric column. Interaction between horizontally adjacent columns are of secondary computational importance as the length scale of the entire troposphere is only about 10km, compared to a common horizontal resolution of about 300km. As schematised in Fig. 2.2 small scale ascendance occurs in grid boxes with deep convective clouds. One approach to calculate this vertical motion would be to calculate
the net mass flux of cumulus cloud into a grid box region for a given temperature and moisture profile. By assuming that to first order, the influx of moisture at low level exits at a higher altitude, the strength of vertical convection would be proportional to the concentration of cumulus cloud within that grid space, with greater deep convection resulting in more clouds.

Figure 2.2: Idealised Schematic of the parameterisation of moist convection: the horizontal fluxes at the top and bottom of the column gives an indirect calculation of the vertical advection required to satisfy mass conservation. The proportion of cloud cover in each grid box is proportional to the strength of vertical motion. Figure taken from Neelin (2011).

2.1.2 Resolution and Cost

The resolution of a model defines the amount of information per unit area it can hold, with higher resolutions having a better representation of a physical quantity to lower ones. A typical representation of ‘low’ and ‘high’ resolution models is shown in Fig. 2.3 illustrating the topographic height of the North American continent. The ‘low’ resolution schematic is on the order of 2° east-west by north-south and is a typical resolution for most models used in the IPCC AR4 report. As a comparison, the ‘high’ resolution schematic utilises 0.2° east-west by north-south. Clearly the latter has a superior representation of the topographical region. The main reason why not all models use such high resolution is primarily due to the exponential increase in computational time (and therefore cost) that comes with enhancing resolution.

Neelin (2011) defines computation time as

\[
\text{Computational time} = (\text{computer time per operation}) \\
\times (\text{operations per equation}) \\
\times (\text{number of equations per grid box})
\]
where an operation can be as simple as a summation or multiplication of two variables. Realistically one needs to take into account other computational limits such as input and output read time, fresh rates etc., making the above equation a very non-conservative estimate.

Computational cost can be roughly scaled as proportional to computational time given the power costs of running a model. Sophisticated models usually need more powerful and expensive computers to run compared to simpler models that are less compute intensive. The absolute cost however becomes irrelevant when the computational time extends way beyond any realistic period of research.

The effects of increasing resolution to computational cost are two fold, affecting both grid box size and the length of each time-step. For the sake of simplicity, say we wanted to half the resolution of a grid box. By doing so we would split all three axis up into two smaller pieces, leading to an increase of $2^3 = 8$ grid boxes to the original one. For computational stability, halving the spatial resolution generally leads to a halving of the time-step (see below). So, for a doubling of resolution, there would be a factor of $2^4 = 16$ increase in the computational time. From the example in Fig. 2.3 the difference in horizontal resolution is a factor of 10, thus a change in the horizontal and temporal axis would already lead to an increase of $10^3 = 1000$ calculations. For simplicity, if we only half the vertical resolution and increase temporal resolution by $10^1$, that still leaves an overall factor of $2^1 \times 10^3 = 2000$, thus the
low resolution model running for 40 years would roughly equal to the high resolution model running for 7 days! With the advent of increasing computational capacity, particularly from virtualisation and Cloud Computing technology where users can ‘rent’ as much computational power as they need, it is becoming faster and easier to run ever more complex models than was previously capable.

The time-step of a model must decrease with increasing resolution to maintain computational stability. For this to be met a criterion must be met known as the Courant-Friedrichs-Lewy (CFL) condition (Laney, 1998), mathematically represented as

$$\frac{u \Delta t}{\Delta x} \leq 1 \quad (2.1)$$

where \(u\) is velocity of the fastest waves in the system, \(\Delta t\) is the time-step length, and \(\Delta x\) is spatial resolution. This states that for a system, the physical domain of dependence must be contained with the numerical domain of dependence. As an example, if the maximum wind speed is \(40 \text{ m s}^{-1}\) and the horizontal grid resolution is \(2^\circ\) or \(\approx 200 \text{ km}\), it would therefore take approximately 83 minutes for a parcel of air to traverse an entire grid box. This would therefore be the largest time-step value available. If the time step value was larger, the parcel would traverse more than 1 grid box and small scale motion would not be resolved. Errors like these can amplify with integration time, leading to spurious errors or even the model ‘blowing up’ i.e. an error message.

Although increasing resolution leads to a better simulation, it does not necessarily mean that it leads to a greater representation of the system being modelled; it may resolve smaller scale processes but the true correction is whether it is modelled correctly. A coarser model with superior parameterisations of small scale processes that has a relatively short spin-up time may in fact be a better model to a slower high resolution model. There is thus always a trade-off between resolution and parameterisation.

2.1.3 Ocean Model

A typical representation of an ocean model is shown in Fig. 2.4 with constant depth as its vertical coordinate. For this example the ocean horizontal resolution is set at 100km. The bathymetry outline (solid black line) models a typical island and a continental land shelf on the right. We note the discretised levels of bathymetry that does not capture all detail but replicates an average over a certain area. Model vertical layer thickness is blown up for the first 500m of the ocean to increase graphical detail. Within this region layer thickness is about 5-10m, increasing to about 50m, enough to resolve the shallow structures of the thermocline.
Below 500m, resolution is much coarser, about 50-250m in thickness, since much of the ocean interior is uniform in properties with small vertical gradients.

![Arbitrary cross-sectional area representing ocean grid resolution. Black continuous curved line represents ‘true’ bathymetry, with the corresponding block representation by the model. Ocean resolution is higher near the surface and becomes progressively coarser the further into the interior. Ocean horizontal spatial resolution is smaller than that of the atmospheric model above. Figure taken from Neelin (2011).](image)

It is computationally more effective for the ocean model to have a higher resolution than its atmospheric counterpart, primarily for two reasons. Firstly, important ocean structures occur at scales far smaller than the resolution of the atmosphere, needing a finer grid spacing to fully resolve these features e.g. the western boundary currents of the Gulf Stream and Kuroshio, and the deep convection region within the North Atlantic (see Chapter 1). Secondly, it is computationally cheaper to run an ocean model that does not have as many calculations per grid box compared to the atmosphere.

Due to the different resolutions of the ocean and atmosphere, the exchange of flux information between the two at the surface interface requires careful interpolation from one grid to the other. Incoming solar radiation is computed in the atmosphere model, being affected by cloud type and cloud concentration. At the surface, LW radiation is exchanged to all levels of the atmosphere, whilst surface heating on the ocean is directly downwards through parameterised mixing including surface wind stress etc. Since sea surface temperature remains relatively stable, this variable is passed onto the atmospheric model where all heat fluxes are calculated, eventually being passed back down to the ocean.
2.1.4 Land and Ice Processes

Land cover equates to roughly one-third of the entire global surface area. Even though property fluxes cannot move through the land as easily as that for the ocean or atmosphere, the presence of land types, vegetation and ice have a direct influence on surface albedo, surface evaporation, sensible heat flux, and therefore the global radiation of the entire system. A land model is a subcomponent of the larger Atmospheric model with fluxes directly interacting with atmospheric properties. Land variables typically involve surface temperature, with soil moisture being a function of precipitation, evapotranspiration rates and vegetation type.

Land surface types are very useful to define since for example, a desert area will have very different properties to that of a rainforest. Some models define a leaf area index (see Stenberg, 2008) which depends on the vertical thickness of vegetation and other parameters linked to soil moisture. These in turn affect the rate of surface and subterranean run-off that eventually find their way back into the ocean. Land vegetation also causes a drag on the wind, slowing down near surface velocities.

Snow cover models may have several layers depending on snowfall, radiation, and near surface temperature and moisture content. Melting and sublimation are calculated from the energy balance between the surface and that of the atmosphere.

Ice models include sea ice fraction to that of open water, sea ice thickness, and velocity. Gaps in between ice, termed ‘leads’, are places where ice melt and surface runoff accumulate into the open ocean. The evolution of ice directly relates to the freshwater flux into the ocean, changing the salinity equations that affect the buoyancy of seawater. Atmospheric fields of wind velocity affect the transport of ice whilst the balance of net incoming radiation depends on ice albedo.

2.1.5 Main Equations used within a Climate Model

The table below lists the main equations with which the majority of climate models are built upon. Some of the listed equations, namely the Temperature Equation, Continuity Equation, Moisture Equation, and Salinity Equation are described below.

The Temperature Equation

The temperature equation is the more common name to the thermodynamic energy equation that involves the time derivative of temperature given changes to a fluid parcel by internal energy, mechanical energy and heating.
For a relatively incompressible fluid such as the ocean, the temperature equation is simply

\[ c_o \frac{d\theta}{dt} = Q \quad (2.2) \]

where \( c_o \approx 4000 \, \text{J} \, \text{kg}^{-1} \, \text{K}^{-1} \) is the heat capacity of sea water, \( \theta \) is potential temperature, and \( Q \) is the net heating rate in \( \text{J} \, \text{kg}^{-1} \, \text{K}^{-1} \). Over the entire vertical length \( H \) of a water column and neglecting lateral fluxes of heat from surrounding areas, the input of fluxes are located at the surface of the ocean, and at the bottom of the ocean. In this way, the temperature equation for the ocean can be re-written as

\[ \frac{d}{dt} \left( \int_{-H}^{0} \rho c_o T \, dz \right) = F_{\text{surf}}^{\text{net}} - F_{\text{bot}}^{\text{net}} \quad (2.3) \]

where the subscripts 'surf' and 'bot' refer to the surface and bottom layers of the ocean respectively, \( \rho \) is density, and the vertical integral is between the surface and the bottom of the ocean \(-H\).

Conversely for the atmosphere, the temperature equation can be written as

\[ c_p \frac{dT}{dt} - \frac{1}{\rho} \frac{d\rho}{dt} = Q \quad (2.4) \]

where \( c_p \) is the specific heat capacity of air at constant pressure. The added term of \( -\frac{1}{\rho} \frac{d\rho}{dt} \) describes the work done on an air parcel through compression and expansion. To understand the effects of changing pressure on temperature we consider a column of air. If we heat an air parcel near the surface it will start to rise. As it rises, the overall atmospheric air pressure decreases due to Hydrostatic balance. The warm air parcel will start to expand going against the surrounding pressure gradient. Expansion of the air parcel is related to its internal energy which in turn, is proportional to its temperature. Thus the expanding air parcel due to increasing altitude will become colder. The reverse will be true if an air parcel descends, and thus contracts, increasing its internal energy and raising its temperature.

**Continuity Equation**

For a closed system of fluid parcels, mass is conserved such that \( dm/dt = 0 \). Density however can change, driving counter-balancing changes to volume. The time rate of change of density can be written as

\[ \frac{d\rho}{dt} = -\rho D \quad (2.5) \]

where \( D = \frac{du}{dx} + \frac{dv}{dy} + \frac{dw}{dz} \) is the 3-dimensional divergence. Divergence and Convergence relate to the spreading out, and coming together of mass. For the atmosphere, pressure effects
are important since air parcels are easily compress, expand, however for the ocean, the non-compressibility of sea-water simplifies calculations.

For the ocean, the continuity equation can be written as

\[ D_o = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = -\frac{\partial w}{\partial z} \]  \hspace{1cm} (2.6)

where \( D_o \) refers to the divergence of the ocean. Eq. 2.6 relates that if changes to density are minute i.e. \( \approx 0 \), we can approximate that a horizontal divergence of fluid must be equal to the ‘feeding’ of fluid by vertical motion into the region.

For the atmosphere, a similar continuity equation exists, written as

\[ D_a = -\frac{\partial w}{\partial p} \]  \hspace{1cm} (2.7)

where \( D_a \) represents the divergence for the atmosphere along isobars, and \( w \) is the vertical velocity in pressure coordinates. Since pressure decreases with height, \( w \) is negative for upward motion. The convergence/divergence of air into/out of a region results in a complementary vertical transport of air, for example, low lying convergence of air at the inter-tropical convergence zone leads to upward motion.

**Moisture Equation**

The water vapour budget in the atmosphere is commonly referred to as the *moisture equation*. It is useful to define the *specific humidity* of air, as the ratio of the mass of water vapour to the total weight of the air parcel. The moisture equation essentially keeps track of sources and sinks of moisture for the atmosphere. If there was no evaporation or precipitation within the air parcel, then moisture would be a conserved quantity, i.e. \( dq/dt = 0 \).

We define the moisture equation as

\[ \frac{dq}{dt} = P_{\text{convection}} + P_{\text{mixing}} \]  \hspace{1cm} (2.8)

\( P_{\text{convection}} \) relates to moisture loss terms such as precipitation, condensation and mixing processes associated with moist convection. \( P_{\text{mixing}} \) includes sources terms of moisture such as surface evaporation and mixing not associated with moist convection. In reality, many processes associated with i.e. Cloud formation, condensation, sublimation etc. happen on unresolvable scales, thus are heavily parameterised.

Over land, the same principles moisture budget principles apply, with the majority of net freshwater input coming from precipitation. Lateral fluxes are now taken into account, for example, from moving river runoff and subterranean flows (depending on the layer complexity of the land/soil model).
2.1.6 Salinity Equation

The salinity of the ocean is dependent on net freshwater flux, with precipitation (evaporation) resulting in a decrease (increase) of ocean salinity within an enclosed region.

We define an idealised situation where a container contains a fixed mass of seawater $M$ and a given volume weighted average salinity $S$; a fixed mass of salt content $m_s = M \times S$. A vertical freshwater flux $F$ exists at the surface as a result of evaporation and precipitation. The time dependent mass equation can be written as

$$\frac{dM}{dt} = -F$$  \hspace{1cm} (2.9)

Using our definition for $m_s$, eq. 2.9 can be expanded into

$$\frac{d}{dt} \left( \frac{m_s}{S} \right) = -F = \frac{1}{S} \frac{dm_s}{dt} + m_s \left( -\frac{1}{S^2} \right) \frac{dS}{dt} = -\frac{M}{S} \frac{dS}{dt}$$  \hspace{1cm} (2.10)

where we have used $dm_s/dt = 0$. Eq. 2.10 can therefore be re-arranged into

$$M \frac{dS}{dt} = SF$$  \hspace{1cm} (2.11)

We define eq. 2.11 as the Salinity Equation. In climate models this is usually approximated further by replacing $M = M_o$ and $S = S_o$ as constant mass and constant salinity respectively. The new $S_o F$ term is termed the “pseudo salt flux”.

The implementation of some of these equations may vary between differing models and complexities but the principle components remain intact. We define equations that are time varying as Prognostic Equations since variables are constantly changing with each time-step; the relevant variable is noted in brackets. The rest are defined as Diagnostic Equation whereby they give the relationship between certain variables. Similar equations like these can thus be combined to eliminate certain variables.

A great many number of climate models do not have to have a fully coupled ocean-atmosphere-land-ice configuration for the purposes of which they are used. One does not have to parameterise subterranean river when analysing incoming solar radiation, as an example. Some are purely aquaplanet models with no land mass (Hayashi and Sumi, 1986; Watanabe, 2005; Enderton and Marshall, 2010) configurations, whilst others only use slab ocean models (Ganopolski, 1998).

2.1.7 Gent-McWilliams parameterisation

Early ocean circulation models such as those used by Bryan (1969) used Laplacian diffusion $\nabla \cdot \nabla \tau$ where $\tau$ is a tracer, in the horizontal and vertical directions to close the equations for
Chapter 2

Summary of Main Equations used by Climate Models

<table>
<thead>
<tr>
<th>Equation name</th>
<th>Model</th>
<th>Notes</th>
<th>Equation Number</th>
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<td>Horizontal Velocity</td>
<td>Atmos/Ocean</td>
<td>Prog.(u,v)</td>
<td>1.10a, 1.10b</td>
</tr>
<tr>
<td>Hydrostatic Equation</td>
<td>Atmos/Ocean</td>
<td>diag.</td>
<td>1.11</td>
</tr>
<tr>
<td>Equation of State</td>
<td>Atmos</td>
<td>Ideal Gas Law</td>
<td>1.13</td>
</tr>
<tr>
<td></td>
<td>Ocean</td>
<td></td>
<td>1.19</td>
</tr>
<tr>
<td>Temperature Equation</td>
<td>Ocean</td>
<td>Prog. (T)</td>
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</tr>
<tr>
<td></td>
<td>Atmos</td>
<td>Prog. (T)</td>
<td>2.4</td>
</tr>
<tr>
<td>Continuity Equation</td>
<td>Atmos</td>
<td></td>
<td>2.6</td>
</tr>
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<td>Ocean</td>
<td></td>
<td>2.7</td>
</tr>
<tr>
<td>Moisture Equation</td>
<td>Atmos</td>
<td>Prog. (q)</td>
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</tr>
<tr>
<td>Salinity equation</td>
<td>Ocean</td>
<td>Prog. (s)</td>
<td>2.11</td>
</tr>
<tr>
<td>Surface pressure</td>
<td>Atmos</td>
<td>1 level</td>
<td></td>
</tr>
<tr>
<td>Surface height</td>
<td>Ocean</td>
<td>1 level</td>
<td></td>
</tr>
<tr>
<td>Surface temperature</td>
<td>Land</td>
<td>1 or more levels</td>
<td></td>
</tr>
<tr>
<td>Soil Moisture</td>
<td>Land</td>
<td>a few levels</td>
<td></td>
</tr>
<tr>
<td>Snow Cover</td>
<td>Land</td>
<td>1 or more levels</td>
<td></td>
</tr>
<tr>
<td>Sea Ice</td>
<td>Ocean</td>
<td>Ice fraction, thickness</td>
<td></td>
</tr>
</tbody>
</table>

Table 2.1: Prog. refers to time varying prognostic equations, with the relevant variable shown in brackets.

Potential temperature and salinity and as an early parameterisation for the effects of mesoscale eddies that could not be spatially resolved (Gent, 2011). It was known 30 years prior that mixing occurred predominantly along isopycnal surfaces rather than across them. The use therefore of horizontal mixing soon proved to have serious disadvantages for ocean modelling. Veronis (1975) showed that horizontal mixing had to be balanced by a false vertical mean velocity. Known as The Veronis Effect, its main consequence was to reduce the northward heat transport of the Atlantic ocean by a short-circuit of its meridional overturning circulation within the sub-tropics (Gent, 2011). Large horizontal diffusion of the order of $10^3 \text{ m}^2\text{s}^{-1}$ implied much stronger cross-isopycnal mixing of the observed order of $10^{-4}\text{ m}^2\text{s}^{-1}$. It was thus agreed that diffusion in z-coordinate models had to be oriented parallel and perpendicular to isopycnal surfaces.

An attempt by Redi (1982) and Cox (1987) to parameterize the adiabatic mixing of eddies for z coordinate models used a mixing tensor $K_{Redi}$ that mixed tracers along constant density...
rather than constant depth surfaces. This mixing for a given tracer $\tau$ is of the form

$$\nabla \cdot \kappa \rho K_{\text{Redi}} \nabla \tau$$

(2.12)

where $\kappa \rho$ is the along isopycnal diffusivity and $K_{\text{Redi}}$ is a tensor that projects $\tau$ onto the isopycnal surface. Since a typical isopycnal slope is of the order of $10^{-2}$ with a maximum of $10^{-2}$ (Gent, 2011), the use of a ‘small slope’ approximation can approximate $K_{\text{Redi}}$ into

$$K_{\text{Redi}} = \begin{pmatrix} 1 & 0 & S_x \\ 0 & 1 & S_y \\ S_x & S_y & |S|^2 \end{pmatrix}$$

(2.13)

where $S_x = -\partial_x / \partial_z \sigma$ and $S_y = -\partial_y / \partial_z \sigma$ are the components of the isopycnal surface.

Gent and McWilliams (1990) expanded on the Redi-Cox scheme by introducing an advection due to eddies, on top of their diffusive effects. This was very important as it allowed for the dynamical behaviour of eddies such as the flattening of fronts (and therefore a destruction of available potential energy) without mixing across isopycnal surfaces. Within the mathematics, the flattening of fronts was completed through the use of “bolus” velocities $\tilde{u}^*$ (see Fig. 2.5), defined by

$$\tilde{u}^* = \begin{pmatrix} u^* \\ v^* \\ w^* \end{pmatrix} = \begin{pmatrix} -\partial_z (\kappa_{GM} S_x) \\ -\partial_z (\kappa_{GM} S_y) \\ -\partial_x (\kappa_{GM} S_x) + \partial_y (\kappa_{GM} S_y) \end{pmatrix}$$

(2.14)

The advection of a tracer from bolus velocities would then equal to

$$\tilde{u}^* \tau = -\kappa K_{GM} \nabla \tau$$

(2.15)

where $K_{GM}$ is represented by

$$K_{GM} = \begin{pmatrix} 0 & 0 & -S_x \\ 0 & 0 & -S_y \\ S_x & S_y & 0 \end{pmatrix}$$

(2.16)

The addition of both the Redi-Cox and GM parameterisations lead to

$$\kappa \rho K_{\text{Redi}} \nabla \tau - \tilde{u}^* \tau = (\kappa \rho K_{\text{Redi}} + \kappa_{GM} K_{GM}) \nabla \tau$$

(2.17)

When $\kappa_{GM} = \kappa \rho$, both Redi and GM matrices led to significant cancellation, resulting in

$$\kappa \rho K_{\text{Redi}} + \kappa_{GM} K_{GM} = \kappa \rho \begin{pmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 2S_x & 2S_y & |S|^2 \end{pmatrix}$$

(2.18)
The complete parameterisation would therefore be of the form

\[ \nabla \cdot \nabla \tau \Rightarrow \nabla \cdot (\kappa_\rho K_{\text{Redi}} + \kappa_{GM} K_{GM}) \nabla \tau \]  

(2.19)

The derived parameterisation is now famously termed ‘GM’ after Gent and McWilliams (1990).

Figure 2.5: Schematic of GM parameterisation: The direction of "bolus" velocities (blue circle) reduces available potential energy by flattening surfaces of constant density without mixing across them. Red arrows indicate along isopycnal diffusion.

2.2 HadCM3 Model

The Hadley Centre couple Model (HadCM3) is the third version of the model developed in the UK. Its primary achievements as stated in Gordon et al. (2000, hereafter G2000) are the absence of flux adjustments. Although a relatively old model by present day standards, HadCM3 is used due to its very good representation of our present climate system but also as a direct comparison to its ‘sister’ model CHIME, discussed later on. Figures that describe HadCM3 are presented in section for CHIME to serve both as direct comparisons between the two models and to avoid duplication of information.

In this next section we describe the model components in detail and the results for components of the climate.

2.2.1 Atmosphere

The atmospheric model HadAM3 is comprised of a horizontal grid spacing of 3.75° east-west, by 2.5° north-south, similar to a spectral resolution of T42. It comprises 19 layers using a hybrid vertical coordinate and has a temporal resolution of 30 minutes.
The improvements from the previous atmospheric model (Johns et al., 1997) include 1. A new radiation scheme which includes 6 (8) different spectral bands in the shortwave (longwave) regions to represent minor trace gases, including \( H_2O, CO_2 \) and \( O_3 \). 2. Additional parameterization of convection on momentum (Gregory et al., 1997) 3. New land surface scheme (Cox et al. 1998) that includes the melting and freezing of soil moisture. The evaporation field includes dependences to vegetation, vapour pressure and \( CO_2 \) concentration. 4. Additional parameterization of orographic drag (Milton and Wilson, 1996) and a new gravity wave drag scheme that includes anisotropy of orography, high drag states and flow blocking, and trapped lee waves (Gregory et al., 1998). 5. The temperature range for mixed cloud phases into ice and water has been tuned from \( 0 - 15^\circ C \) to \( 0 - 9^\circ C \) (Gregory and Morris, 1996) from observational based data (Moss and Johnson, 1994). A parameterisation on cloud droplet size as a function of water content and droplet concentration is included (Martin et al., 1994). 6. Removal of a non-local mixing scheme that affected boundary layer mixing (Smith, 1993).

2.2.2 Ocean

The previous version of the ocean component used an identical grid resolution similar to that of the atmospheric model described above. In the new ocean component the resolution is \( 1.25^\circ C \) by \( 1.25^\circ C \) in the latitude-longitude direction with 20 vertical layers, based on the Cox (1984) model. Vertical resolution is enhanced near the surface to resolve small scale features such as the Thermocline layer, with coarser resolution at depth. Six ocean grid points overlap one atmospheric grid box. Bottom topography was taken from the ETOPO5 (1988) dataset that had a resolution of \( 1/12^\circ \) and interpolated onto the ocean model grid. The Denmark Strait and Iceland-Faeroes ridge was excavated to 7979 m (bottom of level 12) and 534.7m (bottom of level 11) respectively to give a long term outflow of 8.5 \( Sv \), compared to an observed value of 5-6 \( Sv \). An island was placed at the North Pole to remove the singularity formed from converging meridians.

Due to the coarseness of the ocean grid not being able to resolve eddy mixing activity, horizontal eddy mixing of tracers is parameterised using a version of the Gent and McWilliams (GM, 1990) adiabatic thickness diffusion (Visbeck et al., 1997) and the Redi (1982) along-isopycnal-diffusion scheme - full details of GM parameterisation is given in section 3.2.2. Near surface mixing is parameterised using a Kraus-Turner mixed layer sub-model (Kraus and Turner, 1967). Modification of the convective scheme is adjusted for the Denmark Straits and
Iceland-Scotland Ridge to better represent the mixing processes of the overflow waters.

The sea ice model is identical to that used by HadCM2 and includes a simple thermo-
dynamic scheme with parameterisations for ice drift and leads (Cattle and Crossley, 1995). Fractional ice cover is limited to a maximum of 0.995 in the Arctic and 0.980 in the Antarctic since completely unbroken ice is rarely observed. Ice formation happens primarily within the leads, with melting taking place at the surface and at the bottom of the ice. Ice thickness can increase when the weight of snow on top pushes the ice-snow interface below the surface, forming ‘white ice’ (Ledley, 1985).

The sea ice model accounts for changes in salinity due to sea ice changes. Sublimation of sea ice leads to increased salinification of sea water since it is assumed that salt is blown into the leads, with snow fall reducing salinity. Sea ice is assumed to have a salinity concentration of 0.6%, with all rainfall assumed to reach the open ocean through leads.

Surface ice albedo is prescribed at 0.8 at $-10^\circ C$ and below, falling linearly to 0.5 for temperatures between $-10^\circ C$ and $0^\circ C$ as a simple parameterisation for aged snow, surface melt ponds, and bare ice.

Ocean heat flux into the base of the ice is proportional to the temperature difference between the two media, with sea ice temperature assumed to be at $-1.8^\circ C$. A coupling coefficient is parameterised at $20 W m^{-2} K^{-1}$.

Windstress is applied to the ocean through the ice with variables such as ice concentration, thickness and snow depth, being advected within the top layer. To prevent convergence of ice a maximum thickness of 4m is imposed (Steele et al., 1997).

The ocean model uses an Arakawa B grid representation.

### 2.2.3 Ocean-Atmosphere Coupling

The coupling between the ocean and atmospheric models have a frequency of once per day. For each day the atmosphere uses prescribed surface temperature which is used to compute all atmospheric fluxes for each time step. At the end of each day these fluxes are passed back onto the ocean where it is integrated to calculate a new value of SST. This SST again feeds back into the atmosphere and the process repeats itself. For every atmospheric grid cell there are 6 overlapping ocean grid cells overlapping below. To pass information from the ocean to the atmosphere, fluxes are interpolated and averaged onto the atmospheric grid.

Runoff is modelled through river outflow locations at continental boundaries, allowing for interaction with the ocean’s salinity budget. Since rivers are not explicitly modelled, river
runoff is transported instantaneously to coastal areas.

2.3 CHIME Model

The Coupled Hadley-Isopycnic Model Experiment (CHIME) is a GCM currently being developed at the National Oceanography Centre in Southampton, UK. CHIME is in most respects identical to HadCM3 except for the choice of vertical coordinate system used, being isopycnic (or constant density surfaces) instead of constant in depth for the ocean model depth layers. The atmospheric model comprises lateral resolutions of 3.75° east-west by 2.5° north-south (Gordon et al., 2000), with a hybrid vertical coordinate system of 19 layers. Its sea ice model (Cattle and Crossley 1995) is a thermodynamic model that allows for sea ice drift and partial coverage regions for leads (ice free areas). The diffusion and advection schemes of the sea ice model are recoded onto an Arakawa C grid shared with CHIME.

The CHIME ocean model is HYCOM version 2.1.34 (Bleck, 2002). Over the course of the annual cycle the ocean vertical coordinate predominantly follows a purely isopycnal path. For regions where isopycnal surfaces tend to outcrop near the surface, in reality there would be a destruction of certain density layers. To overcome this problem a minimum thickness of 5m is prescribed with the density being allowed to vary. Layer thickness are allowed to expand and contract with a minimum thickness of 5m for the surface layer. For ocean interior layers this minimum thickness increases up 15m by layer 5. The vertical coordinate used is potential density prescribed at a pressure of 2000 dbar, with a thermobaric correction added to the pressure gradient of Sun et al. (1999). Currently there are 25 reference density layers. If the density within a layer changes due to mixing, HYCOM automatically shifts the depth of the layer interfaces such that the density is pushed back towards the reference density of that layer. Although the HYCOM has five extra ocean layers than for HadCM3’s 20 layers, the extra layers are predominantly located near to the surface and so do not substantially change the resolution within the interior ocean. The 25 reference density layers were chosen to reflect the densities of the major oceanic water masses. The increase in resolution towards lower densities is used to distinguish highly mixed water masses near the surface.

The CFL stability criterion (Eq. 2.1) states that for numerical models, the speed of the fastest winds must be less than or equal to the grid spacing divided by the time step (Courant et al., 1928). This ensures that models have an appropriate spatial resolution to fully resolve small scale processes, but more importantly to not ‘blow-up’. Models such as HadCM3 that are based on a spherical grid must use filtering near polar latitudes to avoid diminishing grid
spacing due to convergence of the meridians. This type of filtering is however not appropriate for the HYCOM model where layer thickness must be always finite, and therefore not practical at polar latitudes (Megann et al., 2010). HYCOM therefore employs a spherical bipolar grid (Sun and Bleck, 2001a) with two distinct regions: between 55°N and 78°S the lateral spacing is 1.25° in both zonal and meridional direction (see Fig. 2.6). Northwards of 55°N the grid is bipolar, with the lines of convergence situated in land at locations 110°W and 70°E. The overall effect leads to the removal of the singularity at the geographical North Pole, with increased resolution in the Arctic Region. Under the bipolar region there are exactly six ocean grid boxes under one atmospheric grid box as similar with HadCM3, with a resolution between 40 and 140 km each.

![Figure 2.6: CHIME ocean grid and bathymetry. Regions below 55°N have constant grid spacing whilst above this latitude the grid is bipolar. The poles are located at 70°E and 110°W, marked by red crosses. The different geographical locations of the Bering Strait on the bipolar grid have matching fields for continuity. Figure adapted from Megann et al. (2010).](image)

HYCOM’s bathymetry is derived from Sandwell and Smith (1997), interpolated onto the grid. A minimum depth of 100m is imposed everywhere to prevent numerical barotropic instabilities in shallow water regions (Megann et al., 2010, hereafter M2010). Due to the bipolar grid, the ocean coastlines within the Arctic cannot be identical to that used by HadCM3. Coastlines are prescribed on the ocean model grid, as compared to HadCM3 where they described on the atmosphere. Coastline locations are defined as the zero depth contours after interpolation. These are then adjusted to such that critical Straits are open to realistic depths. The Bering Strait (entrance linking North Pacific and the Arctic) and Gibraltar Strait (entrance linking Mediterranean Sea and Atlantic) are both open and represented by single cell width grid points. Contrasted with HadCM3 that cannot achieve single cell flows due to its ocean B-grid architecture, simulated flow through the Gibraltar Strait is achieved using an exchange algorithm, whilst flow through the Bering Strait remains blocked. Since the coastline northwards of 55°N for the ocean do not match points on the atmosphere due to the bipolar
grid, an interpolation scheme from Sun and Bleck (2001a) is used to conserve fluxes between the two media. Spurious maxima of wind-stress curl are avoided by linearly interpolating onto the ocean grid between adjacent atmospheric cells. Daily coupling of ocean and ice fields to the atmosphere are processed via the OASIS 2.4 coupler (Valcke et al., 2000).

CHIME employs a K-profile parameterisation (KPP) diapycnal mixing scheme (Large et al., 1994) which when used in HYCOM, was found to have substantial improvements when compared to the Kraus-Turner Bulk mixing scheme (Halliwell, 2004).

Sea ice cover was taken from Gloersen et al. (1993) with all sea-ice covered grid cells given an initial prescribed thickness of 2\,m. The model was initialised from the Levitus et al. (2008) Autumn climatologies projected onto the reference densities, and run for 200 years. Greenhouse Gases and Aerosol concentrations are prescribed at pre-industrial levels. The control climate analysis of CHIME by M2010 focused on the time mean period between years 80-119 of spin up, allowing for a direct comparison to HadCM3 which used the same temporal averaging period (Gordon et al., 2000).

2.3.1 Global Mean Diagnostics

Ocean Temperature and Salinity

Both HadCM3 and CHIME were initiated using identical starting conditions taken from the Levitus et al. (1998) climatologies. During the control climate spin-up, global mean top of the atmosphere (TOA) net radiation stabilises by year 60, to a downward magnitude of +0.2 \, Wm\(^{-2}\), implying an initial warming of the ocean. TOA r.m.s. net flux are on the order of 0.3 \, Wm\(^{-2}\), a magnitude similar to HadCM3. The global mean volume weighted potential temperature of the ocean is shown in Fig. 2.7a for HadCM3 and CHIME. Although both ocean models are initiated from identical climatologies, the differences in values at the start of integration are due to the different coastal boundaries and bathymetry depth. The decreasing trend shows CHIME’s ocean cooling by 0.06\,\degree\,C/\,century equal to a global surface heat loss of 0.3 \, Wm\(^{-2}\) for the second century. Compared to HadCM3 that cools by 0.06\,\degree\,C/\,century and is equivalent to a 0.08\,Wm\(^{-2}\) global surface heat loss, the large differences in ocean cooling originates from HYCOM’s inability to conserve heat and salt exactly because of the non-adiabatic time smoothing of layer interfaces in the continuity equation (M2010). Within CHIME this non-conservation amounts to a global mean net surface heat flux of 0.5 \, Wm\(^{-2}\) and a tendency of 0.1\,\degree\,C/\,century, accounting for the differences between TOA and surface radiation imbalance and the calculated ocean cooling rates. Although CHIME exhibits a
non-conservation of heat and salt within the ocean interior, this deficiency is caused more by the time stepping scheme (Leap Frog method in CHIME) rather than exclusively to the chosen vertical coordinate system; the isopynic model GOLD has been shown to conserve heat and salt (Hallberg and Adcroft, 2009). Although CHIME uses a temporal resolution of 36 minutes, a shorter time-step would lead to a reduction in the non-conservation terms, however due to practical integration times, this was deemed unacceptable (M2010). Presently there are developments in CHIME that aims to reduce these issues (A. Megann, 2011, personal communication).

Figure 2.7: Global mean ocean fields for CHIME (solid line) and HadCM3 (dashed line): a). potential temperature, b). salinity, c). sea surface temperature, and d). sea surface salinity. Figure taken from Megann et al., 2010.

Figure 2.7a shows the evolution of global mean sea surface temperature of CHIME and HadCM3. Again the difference in initial values is a result of the different coastal boundaries and bathymetry used between the two models. CHIME exhibits an initial warming for the first twenty years before stabilising for the rest of the integration period. HadCM3 exhibits a general cooling drift in SST showing no signs of stabilisation after 200 years. The warmer SSTs in CHIME compared to HadCM3 are predominantly associated with warm SST values of the Southern Ocean region due to a shallower summer time mixing depth that traps heat near the surface. Although the latter region graphically represents the majority of the SST bias,
there exist other substantially warm areas including continental coastal regions and western boundary currents.

Figure 2.7b shows the evolution of global mean salinity in CHIME and HadCM3. The distribution remains fairly unchanged for the first 80 years but then increases by $0.003\text{psu/century}$, equivalent to a global net evaporation of $2.5\text{mm/y}$, again highlighting the non-conservation of salt within the ocean. For HadCM3 the trend shows minimal changes for the first 60 years, followed by a slight decrease to relatively stable values at the end of integration. The ‘kink’ in the distribution at year 60 represents the imposition of upper and lower limits on the salinity in inland islands (M2010).

The mean sea surface salinity (SSS) is shown in Fig. 2.7d. For CHIME there is a sudden increase in salinity for the first 5 years, mainly originating in the North Atlantic tropical regions, with the trend stabilising over 200 years to similar values initiated at the start of integration. HadCM3 SSS shows a decreasing drift with no sign of stabilisation after year 200.

Figures 2.8a and b shows the SST anomaly for CHIME and HadCM3 respectively from the
Climatology (Josey et al., 1998). The SST error in HadCM3 is less than 0.5°C for the global mean, averaged over the given time period of years 80-119, however there remains significant localised errors. These include a cold bias in the North Pacific of approximately 4°C, cold bias’ between 1 – 2°C within the sub-tropical North Atlantic, and warm bias’ off the eastern coast of the Pacific and Atlantic. The latter has been attributed by G2000 to imprecise parameterisations of marine stratocumulus clouds within the atmospheric model. The Southern Ocean region are generally seen as having a cold bias. The cooling bias within the equatorial region has been attributed to an enhanced Easterly Wind stress (G2000). Whilst no definite argument for the prominent cold bias over the North Pacific is given, the latter authors do describe unrealistic heat fluxes within the region, and a Kuroshio Extension that is too far south compared to real world observations. For CHIME, the eastern coastal warm SST bias’ are due to the same cloud model discrepancy already described with HadCM3. The similar pattern of SST errors southward of South Africa across both models verifies the assumptions of G2000 that it is the low ocean resolution responsible for the observed SST errors. The biggest difference in SST error is the disappearance of the persistent cold bias within the North Pacific. M2010 argues that since subsequent versions of HadCM3, namely, HadGEM1 (Johns et al., 2006), and HiGEM (Shaffrey et al., 2009), both of which use constant depth as its choice of ocean vertical coordinate, exhibit similar cold bias’ for the North Pacific, and therefore due to “some aspect of the dynamics or physics of this model”. CHIME is seen as generally too warm within the North Atlantic and Southern Ocean, the latter region, partly due to the KPP mixing scheme used that produces unrealistic shallow summer mixed layers (M2010).

Figures 2.8c and d show the SSS anomalies for CHIME and HadCM3 respectively, relative to the Levitus et al. (1998) climatologies. Both models share similar errors: the tripolar pattern within the Pacific basin describing a freshening (blue) around 40°N and 20°S, and a salinification (red) within the sub-tropics. The latter signal extends within the sub-tropical Atlantic region, with the South Atlantic being too fresh. For CHIME, the largest differences between models can be observed with an overly salty Arctic region by more than 1 psu with a persistent salinification of the Gulf Stream region. As a comparison, HadCM3 generally has a fresher Arctic region except for pockets of high salinification off the Eurasian and Canadian coasts.

Figure 2.9a and b plots the global mean ocean temperature as a function of depth and time for both models, with values being the differences from the initial state. Both models
exhibit sub-surface warming of about 0.1°C and 1.0°C for HadCM3 and CHIME respectively. This warming trend penetrates into the deeper ocean until a depth of approximately 800m and 500m respectively with CHIME concentrating much of its warming near the surface. Below 800m both models exhibit cooling, though this is small for HadCM3 of about −0.1°C beginning below 1500m. For CHIME there is intense cooling between 600-2000m by about −0.6°C, and for regions below 4000m by −0.3 − 0.2°C. The combined effects of warmer sub-surface layers and colder ‘intermediate’ layers indicates CHIME having a sharper thermocline within the sub-tropics.

The evolution of global mean salinity as a function of depth for the two models is shown in Fig 2.9c and d respectively. CHIME exhibits an initial increase in surface salinity within the top 100m but by the end of integration, stabilises to a value close to initialisation. For HadCM3 there is a persistent freshening near the surface that extends to about 500m at year 90, with a maximum change in salinity of −0.5psu near the surface. Freshening occurs in CHIME, though only at a depth between 100-600m, starting at year 70. The saltier surface waters lie above these fresher waters until the end of integration with no clear mixing between the two. The inherently lower mixing of CHIME may be responsible for this feature, and could suggest a form of ‘layer independence’ between upper surface waters, and those below. Below 600m both models start to increase in salinity, this being more predominant in HadCM3 the
Figure 2.10: Comparisons of meridional ocean heat transport for CHIME (solid) and HadCM3 (dashed). Stars indicate estimates from Ganachaud and Wunsch (2000), whilst dotted lines are from the reanalysis of Trenberth and Caron (2001). Darker shading indicates confidence limits of estimates from NCEP, while lighter shading corresponds to estimates based on ECMWF fluxes. Figure taken from Megann et al. (2010).

Ocean Heat Transport and Overturning Circulation

Figure 2.10 shows the meridional ocean heat transport for CHIME and HadCM3 averaged over the years 80-119 (M2010), along with observational estimates from Ganachaud and Wunsch (2000) and Trenberth and Caron (2001). Generally, both model estimates are within the range of the climatological uncertainty for the majority of latitudes with few exceptions: northwards of 40°N HadCM3 tends to overestimate heat transport, whilst southwards of 55°S CHIME underestimates. An inter-comparison of the heat transport between the two models is described in Chapter 3.

The AMOC streamfunctions for both CHIME and HadCM3 are shown in Fig. 2.11a and b respectively, averaged over the years 80-119. We note that for CHIME, the vertical coordinate has been scaled to height coordinates for a better comparison. Both models take between 60-80 years of integration time before reaching a state of equilibrium. At this stage the maximum AMOC strength is 18-20 Sv (1 Sv = 10⁶ m³ s⁻¹) which at 200 years of integration, remains within 10% of the latter value. The location of the AMOC maximum occurs around 25°N and 50°N at a depth between 800-1000m for both models. The densest overflow waters of CHIME sink to deeper depth that compared with CHIME, along with the pathway of NADW (see Chapter 4 for details). In HadCM3 more than 60% of the downwelling within the subpolar gyre regions is concentrated at 65°N, whereas in CHIME this area is spread out over 10°.
M2010 explains this as CHIME having deeper mixing in the gyre interior, but also to the reduced mixing in the Labrador Sea in HadCM3.

The reverse flow cells of AABW (see Chapter 4 for details) below 4000m is weaker in CHIME than HadCM3, reaching approximately -1 Sv and -4 Sv respectively. These differences have been associated with a spin down of the Antarctic Circumpolar Current (ACC) and the insufficient production of dense waters off the coast of Antarctica due to CHIME having 40% greater annual ice cover than HadCM3. The latter effect shields the ocean from intense heat loss needed for dense water production (Megann et al., 2010).

Figure 2.11: The Atlantic meridional overturning streamfunctions (Sv) for years 80-119 of a). CHIME and b). HadCM3. Light grey shading indicates regions with negative streamfunction values. Figure taken from Megann et al. (2010)

2.4 EPcm Two-Box Model

2.4.1 Introduction

To help better understand the climate system as a whole we use a simple two box model to simulate the Earth system. Developed by Dr Arnaud Czaja of Imperial College, the 2 box Climate Model is named as EPcm and aims at “predicting the time dependent response of Tropical and Extra-Tropical surface temperatures to a given time-dependent change in atmospheric greenhouse gas concentration”. The model is currently in its fourth version of evolution.

2.4.2 Model Variables

The model consists of a tropical ($Eq - 30^\circ N$) and an extra-tropical ($30^\circ N - 90^\circ N$) region. Each has its own box for the atmosphere and the ocean, with the latter being further decomposed into a surface/mixed layer (50$m$ thickness) and a deep layer below representing the thermocline (500$m$ thickness). This configuration represents an Aqua Planet model with no land masses.
Figure 2.12: Model setup of EPcm consisting of two boxed regions for the Tropics and Extra-Tropics. The dashed line represents 30°. The ocean is subdivided into a top mixed layer and bottom thermocline layer. Temperature measurements of the mixed layer are identical those at the surface $T_s$. The model is forced with solar radiation and atmospheric $CO_2$ concentration. There is no seasonal cycle present and only one hemisphere to consider. No land masses are present making an aqua planet representation. The diagram is not to scale.

Only a single hemisphere is being modelled with all variables being driven by the solar energy input and the atmospheric $CO_2$ content.

The average atmospheric temperature of a column of air is computed as

$$T_a = \frac{1}{P_s} \int_o^{P_a} T dP$$  \hspace{1cm} (2.20)

Due to the Aqua Planet representation, the surface temperature is a measure of the average temperature of the ocean mixed layer, computed as

$$T_s = \frac{1}{h_o} \int_{-h_m}^0 T dz$$  \hspace{1cm} (2.21)

where $h_m \approx 50m$ the subscript $s$ represents the surface.

Similarly the average ocean temperature below the mixed layer is computed from

$$T_o = \frac{1}{h_o} \int_{-(h_o+h_m)}^{-h_m} T dz$$  \hspace{1cm} (2.22)

where $h_o \approx 500m$, represents the thickness of the thermocline layer. The subscript $o$ represents the ‘ocean’.
Chapter 2

The ocean and atmospheric heat transports are computed from

\[ H_{O,A} = C_{O,A} \Psi_{O,A} \Delta T_{O,A} \]  

(2.23)

where \( C_{O,A} \) represents the heat capacity, \( \Psi_{O,A} \) measures the strength of the circulation, and \( \Delta T_{O,A} \) represents the temperature difference between the Tropics and Extra-Tropical regions. A more specific form of the equation for \( H_A \) is given in Eq. 3.3. For the atmosphere \( \Delta T \) is the gradient of moist potential temperature, with a trading of warm moist air for cold dry air; in the ocean, \( \Delta T \) is the potential temperature. We note that \( \Psi_O \) is a circulation between the mixed layer and the thermocline layer, characteristic of a wind-driven circulation. No buoyancy driven (Thermohaline) effects are taken into account.

2.4.3 Model Equations

The net top of the atmosphere radiative fluxes for boxes 1 and 2 are computed in \( W m^{-2} \) from

\[ F_T(1,2) = \sigma T_{E(1,2)}^4 - (\epsilon_{(1,2)} \sigma T_{A(1,2)}^4 + (1 - \epsilon_{(1,2)}) \sigma T_{S(1,2)}^4) \]  

(2.24)

where \( T_E \) is the emission temperature seen from space, and \( \sigma \) is the Stefan-Boltzman Constant, \( 5.67 \times 10^{-8} W m^{-2} K^{-4} \). The variable \( \epsilon \) controls the emissivity of the atmosphere due to \( CO_2 \) and \( H_2O \) concentration, and is calculated from

\[ \epsilon_{(1,2)} = 1 - e^{-(\alpha CO_2 + \gamma q_{(1,2)})} \]  

(2.25)

where \( \alpha \) and \( \gamma \) correspond to the emissivity parameters of \( CO_2 \) and \( H_2O \) respectively.

The variable \( q \) corresponds to low level specific humidity and is computed from

\[ q_{(1,2)} = 1000 \times RH_{(1,2)} q_{sat}(\frac{T_{S(1,2)} + T_{A(1,2)}}{2}, 750mb) \]  

(2.26)

where constant relative humidity is assumed at \( RH = 0.6 \). All measurements of \( q \) are calculated at a prescribed level of 750mb.

The turbulent surface fluxes (vertical convection parameterizations) are given by

\[ F_S(1,2) = \Lambda(T_{S(1,2)} - T_{A(1,2)} - \Delta T_Z) \]  

(2.27)

If \( (T_{S(1,2)} - T_{A(1,2)} - \Delta T_Z) > 0 \), \( F_S(1,2) = 0 \) otherwise (Emanuel, 2002). The parameter \( \Lambda >> 1 \) is chosen as a random variable, allowing for a simple representation of “noise” that is
commonly found in measurements of our own climate system.

The atmospheric circulation strength is computed from,

$$\Psi_A = K_A(T_{S1} - T_{S2})$$ (2.28)

where $K_A$ is the circulation strength parameter; the default value being $K_A = 100/15 Sv K^{-1}$. The simple single cell circulation of the atmosphere reflects both Hadley Cell and eddy driven mass transports.

A tight coupling between the atmospheric and ocean circulation strength is designed such that their ratio equals

$$\Psi_O / \Psi_A = 0.1$$ (2.29)

consistent with observational measurements of this measurement near $40^\circ S$ (Czaja and Marshall, 2006). This is a convenient parameterisation since we assume that the ocean circulation is purely wind driven, and therefore changes within the ocean circulation due to the atmospheric wind strength has a relatively fast response time. Although these times are on the decadal scale, particularly at extra-tropical latitudes (Visbeck 1998), the model integration runs are on the order of centuries, smoothing out any transient response features (at least for fixed $CO_2$ concentrations).

Atmospheric heat transport is computed from the product of atmospheric circulation strength $\Psi_A$ and the difference in low level moist static energy $h_A$,

$$H_A = \Psi_A (h_{A1} - h_{A2})$$ (2.30)

where $h_A = c_p T_A + l_v q_A$ is the summation of meridional sensible and latent heat fluxes in the atmosphere. At mid-latitudes, atmospheric heat transport is mostly done by the correlation between the zonal mean differences in meridional velocity and moist static energy ie. $v' h'$. Generally, $h$ includes the geopotential term $\Phi = gz$, where $z$ is vertical height and $g$ is gravitational acceleration. However, the zonal integration of $v' \Phi' = 0$ at mid-latitudes for geostrophic waves. $\Phi$ has thus been eliminated from our equation of $h$.

We additionally define the atmospheric heat transport per unit area as

$$F_A = H_A / \pi R^2$$ (2.31)

The energy conservation equations for the Ocean are given by

$$M_1 \frac{dT_{S1}}{dt} = -F_{S1} + \Psi_O c_O (T_{O1} - T_{S1})$$ (2.32a)
\begin{align}
M_1 \frac{dT_S}{dt} &= -F_S + \Psi_O c_o (T_S - T_{S2}) \tag{2.32b} \\
M_2 \frac{dT_O}{dt} &= \Psi_O c_o (T_{O2} - T_{O1}) \tag{2.32c} \\
M_2 \frac{dT_{O2}}{dt} &= \Psi_O c_o (T_{S2} - T_{O2}) \tag{2.32d}
\end{align}

where \( M_1 = \rho_0 C_0 h_m \pi R^2 \) and \( M_2 = \rho_0 C_0 h_o \pi R^2 \) represent the mass of seawater in the mixed layer and thermocline layer respectively (for a single box). The above four equations are advective, with heat being gained at the surface boxes through \(-F_S\), thus there is no heat generation within the thermocline layer.

The analogous equations to the salinity of the ocean are given by
\begin{align}
M_1 \frac{dS_S}{dt} &= s_o F_o + \Psi_O (S_{O1} - S_{S1}) \tag{2.33a} \\
M_1 \frac{dS_S}{dt} &= -s_o F_o + \Psi_O (S_{S1} - S_{S2}) \tag{2.33b} \\
M_2 \frac{dS_O}{dt} &= \Psi_O (S_{O2} - S_{O1}) \tag{2.33c} \\
M_2 \frac{dS_S}{dt} &= \Psi_O (S_{S2} - S_{O2}) \tag{2.33d}
\end{align}

where \( s_o \) is the mean salinity of the ocean, and \( F_o \) is the freshwater flux at the ocean surface (\( F_o \) is positive leaving the surface (evaporation) and negative into the surface (precipitation)).

Salinity for this model is an inactive tracer and therefore has no impact on \( \Psi_O \). The + and – signs of freshwater flux for the different surface boxes corresponds to evaporation greatly exceeding precipitation within the Tropics, with the reverse being true for the extra-tropics. A salinity gradient can therefore be inferred that is proportional to the freshwater transport in the atmosphere (provided that \( \Psi_O \) and \( s_o \) remain constant). Just as with temperature, salinity is advected from the mixed layers into the thermocline depths.

The energy conservation for the atmosphere is given by
\[ C_A \frac{P_S}{g} \frac{dT_A}{dt} = (F_T + F_S) \mp F_A \tag{2.34} \]

### 2.4.4 Model Diagnostics

#### Hydrological Cycle

The poleward atmospheric moisture transport is computed from
\[ F = \Psi_A (q_{A1} - q_{A2}) \tag{2.35} \]
Since evaporation is given from our convective surface parameterisation scheme, we can infer the precipitation in each box as a residual between the two terms. In steady state, the above equation reduces to

$$|F| = E_{(1,2)} - P_{(1,2)}$$  \hspace{1cm} (2.36)$$

where the modulus sign of $F$ refers to the net evaporation of freshwater in the tropics counterbalancing the net input of precipitation at high latitudes. Freshwater steady state in the atmosphere is reached within 5 years of integration (not shown on figures due to stochastic noise).

To conserve the balance of freshwater transport through a zonal cross-section, the poleward advection of atmospheric moisture must be balance by an equatorward advection of freshwater in the ocean. To calculate this transport we invoke the freshwater salinity budget for the ocean (Stommel and Csanady, 1980).

$$(\Psi_O + F_O) S_O = M (S_O + \Delta S)$$  \hspace{1cm} (2.37)$$

where $F_O$ is the meridional freshwater transport in the ocean (same quantity discussed in the salinity equations), $S_O$ is the mean ocean salinity, and $\Delta S$ is the salinity difference between poleward and equatorward moving waters. The above equation can be re-arranged such that

$$F_O = \Psi_O \frac{\Delta S}{S_O}$$  \hspace{1cm} (2.38)$$

Oceanic freshwater transport can be computed as dependent on the oceanic circulation strength, and the salinity gradient. We label Eq. 2.38 as the *indirect* calculation of $F$.

As discussed in Chapter 1, the assumption of fixed relative humidity ensures that the Clausius Clapeyron relationship is strictly obeyed, with low level values of $q$ increasing at 7%/K; precipitation (and in this case, precipitation) however does not does have to follow this rule. To understand this relationship further we employ Eq. (5.2). Freshwater transport would indeed increase with C-C scaling providing that atmospheric mass transport $\Psi_A$ remains constant. As we shall see later this is not the case, with the latter term in fact decreasing, though at a rate smaller than C-C. From the product changes of the R.H.S terms of the latter equation, freshwater transport thus increases with increasing surface temperatures but at a rate smaller than C-C scaling.
Vertical Ocean Heat Transport

By summing the time evolution equations of surface temperature we get an expression for the ocean heat content for the upper ocean,

$$\frac{d}{dt} \left[ \rho_{OC} h_m (T_{S1} + T_{S2}) \right] = - (F_{S1} + F_{S2}) + \frac{\Psi_{OC}}{\pi R^2} (T_{O1} - T_{S2})$$ \hspace{1cm} (2.39)

Conversely the heat content for the thermocline layer is

$$\frac{d}{dt} \left[ \rho_{OC} h_m (T_{O1} + T_{O2}) \right] = - \frac{\Psi_{OC}}{\pi R^2} (T_{O1} - T_{S2})$$ \hspace{1cm} (2.40)

The repeating term $Q_O = \frac{\Psi_{OC}}{\pi R^2} (T_{O1} - T_{S2})$ thus is a measure of the vertical heat flux between the upper and deep ocean; total heat content does not change but is merely redistributed within the vertical.

Relating the T/S curve to the ratio of Ocean to atmosphere Heat Transport

Just as $\Delta S$ referred to the difference in salinity between mass weighted poleward and equatorward moving water masses, so we can define $\Delta T$ as a similar quantity for temperature.

We infer the Dry Static Energy transport of the atmosphere (DSE) as the difference between the total atmospheric heat transport and its latent heat component. We further define the ratio of DSE to latent heat transport $L$ as

$$\gamma = \frac{DSE}{l_v F}$$ \hspace{1cm} (2.41)

where we have rewritten latent heat transport as the product of freshwater transport and the enthalpy of vaporisation.

Through careful manipulation of the salinity budget equations and the above expressions we can relate the ratio of ocean to atmospheric heat transport as proportional to the ratio of $\Delta T/\Delta S$, and inversely proportional to $\gamma$. The in-depth details of the theory are included in Chapter 4 and are not described in detail here to eliminate repetitiveness.

Diagnostics for this new ratio of $H_O/H_A$ is termed as an indirect calculation due to its dependency on $\Delta T$ and $\Delta S$; indirect meaning that the latter quantities need to reach steady state for the solution to convergence to the correct answer.

2.4.5 Model Control Parameters

2.4.6 Control Climate Results

Due to the model’s ‘tuning’, the control climate produces realistic magnitudes of quantities similar to that of our present day climate. The natural variability of the system is much weaker.
Table 2.2: Default model parameter values for the EPcm Two-Box model.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thickness of Mixed Layer</td>
<td>$h_m = 50m$</td>
</tr>
<tr>
<td>Thickness of Thermocline layer</td>
<td>$h_O = 500m$</td>
</tr>
<tr>
<td>Low Level relative humidity (Box 1 and 2)</td>
<td>$RH_1 = RH_2 = 0.6$</td>
</tr>
<tr>
<td>Effective Heat Capacity</td>
<td>$C_A = 2000 J kg^{-1} K^{-1}$</td>
</tr>
<tr>
<td>Critical vertical temperature gradient</td>
<td>$\Delta T_Z = 40K$</td>
</tr>
<tr>
<td>Circulation Strength Parameter</td>
<td>$K_A = 100/15 S v K^{-1}$</td>
</tr>
<tr>
<td>Emissivity parameter for water vapour</td>
<td>$\gamma = 1.25$</td>
</tr>
<tr>
<td>Emissivity parameter for carbon dioxide</td>
<td>$\alpha = 1.2 \times 10^{-3}$</td>
</tr>
<tr>
<td>Carbon dioxide concentration</td>
<td>$CO_2 = 280 ppm$</td>
</tr>
<tr>
<td>Ratio of Mass Transport</td>
<td>$\Psi_O/\Psi_A = 0.1$</td>
</tr>
<tr>
<td>Emission Temperature (Box 1)</td>
<td>$T_{E1} = 268K$</td>
</tr>
<tr>
<td>Emission Temperature (Box 2)</td>
<td>$T_{E2} = 240K$</td>
</tr>
<tr>
<td>Large number for convective parameterizations</td>
<td>$\Lambda = 100 \times [1 + 0.05 \varsigma(t)]$</td>
</tr>
</tbody>
</table>

than that of the real world due to the very “static” dynamics used to parameterize the ocean and atmospheric circulations and the lack of a seasonal cycle.

The warmest ocean temperature is found within the Tropical surface mixed layer depth (Fig. 2.13a, holding a consistent value of 299.9K). In steady state $T_{S2} = T_{O1} = T_{O2}$ (Fig. 2.13b); this solution can easily be reached from Eqs. 2.32c and 2.32d whereby in steady state, the rate of change of temperature is zero.

The ocean temperature of the Thermocline layer originates at the extra-tropical surface before it is subducted into the interior. A global mean temperature of 290.3K (Fig. 2.13c) is reached, consistent with present day estimates from global satellite systems; a meridional
surface temperature gradient is inferred as the difference between warm tropical and cooler extra-tropical waters, calculated at $19.2^\circ C$ (Fig. 2.13d).

At equilibrium the mean TOA radiation is zero, however daily fluctuations are on the order of $0.03 \text{ Wm}^{-2}$ (Fig. 2.14a). During spin up of the model the surface mixed layers warm the fastest reaching a steady state temperature after twenty years (Fig. 2.14b, red and blue curves). The Thermocline layer warms up much slower, with excess heat being stored at this depth after equilibrium temperatures have been reached with the surface layers (cyan and magenta curves). Steady state heat storage between the thermocline tropics and extra-tropical regions requires an integration time of 1000 years or more. This longer time period in comparison with the mixed layer is much longer due to the increased layer thickness.

Net surface heat fluxes for box 1 and 2 are $-7.7 \text{ Wm}^{-2}$ and $7.7 \text{ Wm}^{-2}$ respectively with the sign convention being negative (positive) values meaning net absorption (emittance) of radiation (Fig. 2.14c). The regional sum of each region represents the aforementioned global net TOA radiation of zero. Evaporative cooling for box 1 remains approximately $80.1 \text{ Wm}^{-2}$ as outgoing longwave radiation (Fig. 2.14d). Box 2 exhibits no evaporative cooling for the control climate due to the relatively cold surface temperatures. We shall see later on that with higher $CO_2$ concentrations and therefore higher surface temperatures, convection develops in box 2.

Salinity is initialised at 35 psu across the whole ocean. It is noted that a minority of the coming figures have high resolution scales to properly show the stochastic nature of the model. For the tropical surface salinity, stabilisation occurs after 50 years at a value $\approx 36.3 \text{ psu}$ for the tropical mixed layer (Fig. 2.15a, red curve). The salinity of the thermohaline layer takes roughly 300 years to reach salinity equilibrium with its surroundings. As the governing equations of salinity are highly similar to that of temperature, $S_{S2} = S_{O1} = S_{O2} \approx 34.9 \text{ psu}$ (Fig. 2.15b). The large dip in salinity values for the extra-tropical thermocline layer at around 10 years of integration (Fig. 2.15b, blue curve) originate from the lag time between the high tropical surface waters obtaining a high enough salinity value to be passed into the deeper layers. The salinity ‘minimum’ however recovers and after 100 years, the salinity fields of $S_{S2}$, $S_{O1}$, and $S_{O2}$, start to converge as the ocean salinity distribution reaches equilibrium with that of the atmosphere. The equality between the thermocline regions and the extra-tropical surface layer must equal to each other at equilibrium to the salinity equations. From Eq.

Both ocean and atmospheric heat transports (Fig 2.16a) gives comparable estimates to real world observations of approximately $1.6 \text{ PW}$ and $3.8 \text{ PW}$ at $30N$ respectively, with a total
Figure 2.13: Model temperature evolution with time: Box 1 tropical region for mixed layer (red) and thermocline (magenta) a). Box 2 extra-tropical region for mixed layer (blue), thermocline (cyan), with inserted tropical thermocline layer (magenta) for comparison b). Global mean surface temperature c). and surface equator to pole temperature gradient d).

Figure 2.14: Global TOA net radiation centred at 0Wm$^{-2}$ with daily fluctuations of $\pm0.03Wm^{-2}$ a). Heat storage evolution for each ocean region b). Net surface heat fluxes with negative (positive) values constituting heat into (out) of surface c). Evaporative cooling for each region. The tropics are highly evaporative whilst extra-tropics have no evaporation d)
Figure 2.15: Model salinity evolution with time: Box 1 tropical region for mixed layer (red) and thermocline (magenta) a). Box 2 extra-tropical region for mixed layer (blue), thermocline (cyan), with inserted tropical thermocline layer (magenta) for comparison b). Global mean ocean salinity c). and surface equator to pole salinity gradient d).

Figure 2.16: Meridional heat transport for the ocean (cyan) and atmosphere (magenta) a), with total (ocean and atmosphere) system heat transport b). Atmosphere (red) and ocean (blue) circulation strength, with the ratio of the two equal to 0.1, consistent with model parameterization c). Rate of precipitation for the tropical (red) and extra-tropical (blue) region d).

heat transport magnitude close to $5.5PW$ (Trenberth and Caron, 2001). The over estimation of $H_O = 0.98PW$ is due to the aquaplanet configuration of the model, with a much larger
Figure 2.17: Model atmospheric CO₂ concentration fixed at 280 ppm. a). Lower tropospheric specific humidity transport between each region. b). Surface and atmospheric temperature difference between each region. c). Atmospheric heat transport (magenta) decomposed into its Dry Static Energy (blue) and Latent (red) heat components.

Figure 2.18: Meridional freshwater transport calculated directly in the atmosphere (red) and indirectly (green) from salt conservation arguments. a). The ratio of ocean to atmospheric heat transport calculated directly (blue) and indirectly from salt conservation theory (red) b). The ratio of DSE to Latent heat transport in the atmosphere a.k.a. γ c). The ratio of $\Delta T/\Delta S$ a.k.a. $\Gamma$ and the T/S slope. d). We note the convergence of indirect with direct methods of calculations in a). and b). as the ocean comes into equilibrium with the atmosphere.
ocean being able to transport more heat. The lower calculated value of atmospheric heat transport \( H_A = 3.54 \text{PW} \) is surprisingly similar to the observed value, given the simple 2 box configuration of the model. Total heat transport (Fig. 2.16b) has a value of 4.52\text{PW}; although smaller than that observed, it is still similar in range to a wide range of extra-tropical latitudes.

Oceanic mass transport is exactly an order of magnitude smaller than for the atmosphere highlighting the instantaneous coupling between the two media via the ‘wind-driven’ circulation architecture (Fig. 2.16c), and consistent with model parameterisations from (2.29).

The precipitation rate within the tropical region is approximately 7 times that within the extra-tropics (Fig. 2.16d). Incidentally the precipitation rate for the latter region is a measure of strength for the hydrological cycle, since this moisture must be transported polewards from the Tropics.

Atmospheric concentrations of \( \text{CO}_2 \) are kept constant at 280\text{ppm} throughout the control integration (Fig. 2.17a). The regional contrast between lower tropospheric values of specific humidity is \( \approx 3.71 \text{g/kg} \) consistent with the Tropics having a more moist atmosphere (section 2.4.4).

The area mean atmospheric temperature contrast remains similar, albeit slightly cooler than surface temperatures contrasts by \( \approx 17.5^\circ \text{C} \) (Fig. 2.17c).

Latent heat transport in the atmosphere (the product of freshwater transport and enthalpy of vaporization \( l_v \)) is computed at approximately \( L = 1.2 \text{PW} \) (Fig. 2.17d, red curve). Comparing to NCEP derived observations of \( L \) at 40\text{°N(S)} of approximately 1.8(1.5)\text{PW} (Pierrehumbert, 2002), the model under represents latent heat transport by about one third. One would assume that latent heat transport would be over represented within an aqua planet model, since the whole of the tropical region is a source of freshwater. Freshwater transport however depends on both \( \Psi_A \) and \( \Delta q \). Assuming that the specific humidity gradient is representative of our present climate (temperatures of the column integrated air temperature are well reproduced), we can conclude that \( \Psi_A \) has a smaller value; a result that may be an effect of the simple diffusive parameterisation used in computing \( K_A \). This would be consistent with the reduced value of \( H_A \) that is also proportional to \( K_A \).

Dry static energy transport dominates the atmospheric heat composition (Fig. 2.17d, blue curve) with its value being almost twice that of its latent heat counterpart at 2.4\text{PW}. Compared with real world observation of 2.5 (2.6) \text{PW} at 30\text{°N(S)} respectively, (Pierrehumbert, 2002), EPcm represents DSE surprisingly well.

The following sections of analysis are largely applicable to theoretical variables described
in Chapter 4 for the applications of the T/S curve.

The model’s value of $\gamma$ (DSE/L) is computed at $\gamma = 2.0$, compared to observations from the previous numbers given of $\gamma = 1.4$ (1.7) at 30° N (S) respectively. Although quite larger, the model’s value is of similar magnitude and very nearly equal to real world observations and other model calculations at higher extra-tropical latitudes (Pierrehumbert, 2002; Chapter 4).

Direct model measurements of atmospheric moisture transport and that inferred from $F = \Psi_O \Delta S/\Delta S_O$ (Fig. 2.18a) graphically illustrate the results of the two methods converging with time due to the salinity distribution reaching equilibrium with surface freshwater flux. This is also mirrored with the ratio of ocean to atmospheric heat transport (Fig. 2.18b) from direct and inferred theory that is also based on salinity equilibrium arguments.

Calculations of the ratio of $H_O/H_A$ are $\approx 0.3$ (Fig. 2.18b), showing both the direct and indirect calculations discussed previously. Model results are slightly higher than observed values (Czaja and Marshall, 2006, Fig. 10), consistent with the model’s overestimation (underestimation) of $H_O$ ($H_A$) due to the aqua planet geometry discussed previously.

The T/S slope value is calculated at $\Delta T/\Delta S = 14.8K/psu$ (Fig. 2.18d), and is similar to readings from the HadCM3 model (see Chapter 4).

A summary of the control climate results is shown in Table 2.3.

### 2.5 Model Runs and Scenarios

In this section we describe the details of all the different CO$_2$ scenarios used in this report.

#### 2.5.1 Control Scenarios

We define a Control climate scenario as one where the model has been run using a prescribed atmospheric CO$_2$ value of 280 ppm, similar to pre-industrial conditions. For both HadCM3 and CHIME.

For the HadCM3 and CHIME no explicit integrations were made, with all used data taken from archived runs. For HadCM3, all data was downloaded from the British Atmospheric Data Centre website (www.badc.ac.uk) whilst for CHIME, data was downloaded from the National Oceanography Centre’s internal server (www.noc.soton.ac.uk).

We define the control climate scenarios for both HadCM3 and CHIME as being a 40 year period between the years 80-119 of spin-up; For HadCM3 this relates to the AAXZCA Control run, whilst CHIME’s run ID is identified by the letters ‘cD’. Both data sets have monthly
## Chapter 2

### Control Climate Numbers

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface temperature of Tropics</td>
<td>$T_{S1} = 299.89K$</td>
</tr>
<tr>
<td>Surface temperature of Extra-Tropics</td>
<td>$T_{S2} = 280.69K$</td>
</tr>
<tr>
<td>Global surface temperature</td>
<td>$T_S = 290.29K$</td>
</tr>
<tr>
<td>Thermocline temperature (Tropics)</td>
<td>$T_{O1} = T_{S2}$</td>
</tr>
<tr>
<td>Thermocline temperature (extra-Tropics)</td>
<td>$T_{O2} = T_{O1} = T_{S2}$</td>
</tr>
<tr>
<td>Surface salinity of Tropics</td>
<td>$S_{S1} = 36.3psu$</td>
</tr>
<tr>
<td>Surface salinity of Extra-Tropics</td>
<td>$S_{S2} = 34.9psu$</td>
</tr>
<tr>
<td>Thermocline salinity (Tropics)</td>
<td>$S_{O1} = S_{S2}$</td>
</tr>
<tr>
<td>Thermocline salinity (extra-Tropics)</td>
<td>$S_{O2} = S_{O1} = S_{S2}$</td>
</tr>
<tr>
<td>Mean Ocean Salinity</td>
<td>$S_O = 35psu$</td>
</tr>
<tr>
<td>T/S Slope</td>
<td>$\Gamma = \Delta T/\Delta S = 15K/psu$</td>
</tr>
<tr>
<td>Low level moisture of Tropics</td>
<td>$q_1 = 4.78g/kg$</td>
</tr>
<tr>
<td>Low level moisture of Extra-Tropics</td>
<td>$q_2 = 1.07g/kg$</td>
</tr>
<tr>
<td>Atmospheric circulation strength</td>
<td>$\Psi_A = 128 \times 10^9kgs^{-1}$</td>
</tr>
<tr>
<td>Atmospheric moisture transport</td>
<td>$F=0.47 \times 10^9kgs^{-1}$</td>
</tr>
<tr>
<td>Atmospheric heat transport</td>
<td>$H_A = 3.54PW$</td>
</tr>
<tr>
<td>Oceanic heat transport</td>
<td>$H_O = 0.98PW$</td>
</tr>
<tr>
<td>Total Heat transport</td>
<td>$H_A + H_O = 4.52PW$</td>
</tr>
<tr>
<td>Ratio of Ocean to Atmospheric Heat Transport</td>
<td>$H_O/H_A = 0.3$</td>
</tr>
<tr>
<td>Latent Heat Transport L</td>
<td>$L = 1.2PW$</td>
</tr>
<tr>
<td>Dry Static Energy Transport DSE</td>
<td>$DSE = 2.4PW$</td>
</tr>
<tr>
<td>Ratio of DSE/L</td>
<td>$\gamma = 2.0$</td>
</tr>
<tr>
<td>Hydrological cycle</td>
<td>$E=40\pm2.3Wm^{-2}$</td>
</tr>
</tbody>
</table>

Table 2.3: Model variable results under control climate settings for the EPcm Two-Box model.

Mean temporal resolution. All calculations were done for each month for the entire 40 years and then averaged over $12 \times 40 = 480$ times. These data sets were widely used for Chapter 3, which compared the control run scenarios of HadCM3 and CHIME.

For the EPcm model, the control climate refers to the first 300 years of integration at pre-industrial $CO_2$ levels.
2.5.2 Transient Scenarios

A transient scenario deals with how the climate changes when forced by a temporally changing concentration of atmospheric $CO_2$. For CHIME, a transient scenario existed whereby the model was run at pre-industrial $CO_2$ concentrations for 60 years, whereby afterwards $CO_2$ increased by 1% per year up to the end of integration at year 200; $CO_2$ concentration at this stage was quadruple pre-industrial values. As a note, this scenario has the model ID run ‘cG’.

For CHIME, we defined 4 distinct time frames, known as Transient Scenarios. Each period was 10 years in length and defined by the years 85-94, years 125-134, years 165-174, and years 190-199, corresponding to a time mean $CO_2$ concentration of 1.4, 2.0, 3.0 and 4.0 times pre-industrial conditions. The specified dates are centred on the stated $CO_2$ concentration e.g. $2 \times CO_2 = 560 ppm$ is reached at year 170 and not 165. As such, the stated $4 \times CO_2$ period actually averages to only 3.8 times, however we round this value up to 4.0 for convenience. The scenario with $1.4 \times CO_2$ represents present day values of approximately 380 ppm. The reference of this scenario will be described by its multiple, followed by the letters TRANS eg. 3 times $CO_2$ period would be named as $3 \times TRANS$.

We do not use any transient scenarios for HadCM3.

For EPcm we describe two distinct methods of explicitly running the model with different $CO_2$ values. The first is a fixed $CO_2$ forcing, whilst describes the classical transient scenario of a steady rate of increase in $CO_2$.

For the fixed $CO_2$ forcing scenarios, the model is forced with instantaneous multiples of pre-industrial $CO_2$ concentration at the start of integration, with the values 1.0, 2.0, 3.0 and 4.0, and run for 300 years.

The transient scenario starts off at 280 ppm and increases at a rate of 1% per year until year 140 (similar to CHIME). The model is then further integrated for another 160 years at a fixed concentration of 1120 ppm (4 times pre-industrial value) to stabilise. Thus, both fixed and transient scenarios have 300 years of integration time apiece.

2.5.3 Anomalous Scenarios

Anomalous scenarios describes the differences between climate states at higher $CO_2$ concentrations from that run at pre-industrial conditions. Hence, for CHIME, there will be 4 distinct anomalous climate scenarios relating to the 4 different transient scenarios.

Anomalous scenarios are used extensively Chapters 5 and 6 where we look at changes in Heat Transport and the Hydrological Cycle in CHIME.
2.6 Summary

The construction and evolution of climate models have been shown to depend heavily on the technological advancements of computational power. Depending on the resources at hand to any research team, there is usually a compromise between resolution and parameterisation vs. computational time and cost. The pros and cons must be weighed out since increases in resolution at a longer integration time may just be as good as better parameterisations at faster times. What is certain is that computational power will increase and become cheaper as developments in the field evolve, prompting higher resolutions in any situation.

Our three main models used in this study are the large GCMs of HadCM3 and CHIME, and a two box model named EPcm. The former two are almost identical model except for a change in ocean vertical coordinate and corresponding ocean mixing scheme. Whilst HadCM3 uses constant depth in the z-direction (Bulk Formula mixing scheme), CHIME employs an isopycnic coordinate system (KPP mixing scheme). The comparison of the two models is thus a perfect test bed for observing how changes in the ocean or atmospheric system is affected simply by the change in vertical coordinate system.

Both models represent present day climatologies fairly well. Major anomalous features between the two remain, with HadCM3 described as having an overly cold SST bias, whilst CHIME exhibits warm SST anomalies. Conversely for salinity, HadCM3 is observed to be much too fresh, with the opposite in CHIME, being overtly saline near the surface.

For its simplicity, the EPcm model reproduces many features of the present climate at mid-latitudes, including heat transport, freshwater transport, temperature and specific humidity. The model is simple enough to understand the mechanics of the climate which will be used to explain important conceptual responses of the climate system to anthropogenic forcing in later chapters.
CHAPTER 3

Meridional Heat Transport and the Hydrological Cycle

3.1 Introduction

The distribution of heat around our planet is an essential process for maintaining our present climate state. Its advection is carried by the oceans and atmosphere from the warm tropics to the colder extra-tropical regions, resulting in a decrease of the overall equator-pole temperature gradient. Without it, the extremes in surface temperature we find near the equator and poles would be greater, and possibly limiting the global surface area where life would thrive. Understanding heat transport is thus a very important part in understanding our climate.

The transport of heat comes in many different forms, one being latent heat, which is the transport of moisture or freshwater. The transport of moisture in the atmosphere has different latitudes of convergence and divergence, resulting in net evaporation and precipitation at the surface.

In the ocean, freshwater transport is based on the net input and output of moisture at the surface, affecting the ocean’s salinity that in turn is responsible for buoyancy forcing, particularly at high latitudes. Such changes to the ocean’s stratification affect the vertical transport of seawater that in some areas can be seen as a proxy for the strength of the overall ocean circulation.
In this chapter we describe and discuss the basics of heat transport, beginning with an overview of the subject for the last 60 years (section 3.2.). We then go on to discuss the main components of heat transport with specific references to the differences between the time varying and time mean components, and spatially varying and spatially mean components (section 3.4).

The main body of work pertaining to heat transport under a control climate scenario is split into two parts: the first dealing with heat transport in the EPcm model and its sensitivity to changing certain parameters (section 3.5), and secondly a comparison of model heat transports for CHIME and HadCM3 (section 3.6).

The Hydrological Cycle is discussed in section 3.7 starting with past observations in the field. The main body of work for this subject compares the hydrological cycle of HadCM3 and CHIME under a control climate scenario, concentrating predominantly on model differences in net freshwater flux for the major ocean basins (section 3.8).

### 3.2 History of Heat Transport

#### 3.2.1 Observations

Some of the earliest works on the Earth’s radiation budget were conducted by Houghton (1954), researching the annual variations of the Northern Hemisphere’s heat budget through their present day observed data on surface radiation and cloud albedo measurements.

By the 1960s Lorenz (1967) had published a general theory on the atmospheric circulation. For the ocean, theorist such as Stommel (1948) and Sverdrup (1957) were laying the foundations on ocean circulation. Empirical evidence at that time suggested that the heat transport by ocean currents was the same order of magnitude as that of the poleward energy transport in the atmosphere (Sverdrup, 1957). Attempts to measure the oceanic mass transport were attempted by Bryan (1962) using hydrographic data, however due to the sparseness of spatial data coverage of net radiation gain and loss at the ocean surface, this method proved difficult to ascertain clear answers. Vonder Haar and Oort (1973) attacked this problem at a different angle, using new satellite data on the net radiation budget of the atmosphere system to compute a time mean value of the top of the atmosphere (TOA) energy budget. Results during this time period suggested an almost equal magnitude of heat transport for both the ocean and atmosphere of approximately 2\(PW\). We note that the error bars of the ocean are as large 1\(PW\) deviation from annual mean measurements. Oort and Vonder Haar (1976) expanded on this...
research, this time looking at the annual variations of heat flux for the Northern Hemisphere, in particular the temporal change in seasonal heat storage of the ocean which was found to transport a maximum heat flux of $-8\text{PW}$ in August.

The question of why the total poleward heat transport is remarkably anti-symmetric about the equator given the different hemispheric heat transport distributions of the ocean and atmosphere was tackled by Stone (1978) who showed that the total hemispheric poleward heat transport could be completed to a fair degree of accuracy, by just the solar constant, the radius of Earth, the axial tilt, and the hemispheric mean albedo; assuming that the hemisphere was in equilibrium and that ocean-atmospheric structures were dominated by the planetary scale. His work suggested that underlying features such as land configurations and orography were insensitive to the total heat transport, highlighting the tight coupling between the ocean-atmospheric system as a whole, rather than separate entities. Stone’s dynamical framework tied in with work of Bjerknes (1964), stating that changes in either the ocean or atmospheric heat transport tend to be compensated by the other for the total heat transport to remain constant.

In the early 80s research by Stone and Miller (1980) showed that the total atmospheric heat transport was well correlated with the zonal mean latitudinal temperature gradient, in particular the combined eddy flux (transient and stationary parts combined, see section 3.4.1.) at mid-latitudes. The ocean however showed no correlation between heat flux and temperature gradient at low latitudes.

Work by Stommel and Csannady (1980) laid the mathematical framework for linking freshwater transport in the atmosphere and the oceanic heat transport to temperature and salinity distributions in the ocean. Their theory assumed an equilibrium state for both the ocean and atmosphere such that the freshwater transport was balanced by the salinity distribution in the ocean.

With the advent of better hydrographic data sets of the ocean, the emphasis of measuring the ocean heat transport shifted from the indirect air-sea flux balance calculations of Budyko (1963) to more direct calculations using heat content and mass transport of the ocean. The seminal work by Hall and Bryden (1982) calculated the direct heat transport of the Atlantic Ocean at $25^\circ N$ to be $1.2 \pm 0.3\text{PW}$; far more accurate than their present day indirect methods (see section 3.3.1). As with the Atlantic and due to the coarseness of data sets available, regional studies of the ocean became common, such as for the Southern Ocean (De Szoke and Levine, 1981; Baker, 1980; Gordon and Molinelli, 1981) and the North Atlantic (Hall and
Bryden, 1982; Wunsch, 1980) where the meso-scale eddy heat fluxes were seen as an important mechanism for heat transport.

By the mid-80s enough oceanic data was available both spatially and temporally to produce the first global oceanic heat transport representation, decomposed into individual basin parts (Hastenrath, 1982; Hsiung, 1985). Using combined data from surface marine observations spanning 1949-1979, Hsiung (1985) used a surface-energy balance model to compute the latitudinal distribution of heat transport for each basin. The results inferred that the Atlantic basin transports heat northwards even in the Southern Hemisphere, the Indian basin has a southward heat transport, whilst the Pacific basin transported heat polewards in both hemispheres. From sensitivity studies it was found that the latter region had the largest uncertainties in both magnitude and direction, whilst the former regions were stable in their measurements.

Although a global coverage of surface heat fluxes were attainable, the magnitudes of ocean heat transport were still somewhat inconsistent, with research by Oort and Vonder Haar (1976) and Carissimo et al. (1985) having estimates of between $3 - 3.5 \, PW$ at $20^\circ N$, whilst Hastenrath (1982), Talley (1984) and Hsiung (1985) estimated ocean heat transport to be between $1 - 2 \, PW$. Masuda (1988) explains that due to the sparseness of data in the Tropics and the Southern Ocean, this may have led to an overestimation (underestimation) of oceanic (atmospheric) heat transport.

In conjunction with present day measurements of the climate system, paleo-scientists were starting to look at past climate events of the inter-glacial periods from $CO_2$ proxies in ice cores, and how this would have affected the climate system (Boyle and Keigwin 1986; Broecker et al. 1985; Broecker and Peng 1986; Broecker and Peng 1987; Broecker 1987).

During the mid-90s the assimilation of all current atmospheric data was used for energy balance computations and adjusted to fit physical constraints. These ‘reanalysis’ data sets aimed to reduce the errors associated with measuring heat transport through the utilisation of the latest satellite data. The two main reanalysis data sets were from the National Centers for Environmental Prediction (NCEP; Kalnay et al. 1996) and the European Centre for Mid-Range Weather Forecasts (ECMWF; Gibson et al. 1997).

Trenberth and Caron (2001), hereafter TC01, made use of both reanalysis data sets to calculate the atmospheric heat transport; data from the ERBE satellite that was launched in 1984, was used to calculate the total meridional heat transport. By calculating the ocean heat transport as a residual between the total and ocean heat components, and also from
the latest ocean climate models at that time, TC01 further showed increases in the poleward heat transport for the atmosphere when compared to previous estimates. At peak values, atmospheric heat transport was estimated at 5.0\( \text{PW} \) at 43 \( ^\circ\text{N} \), compared to 3.1\( \text{PW} \) (Oort and Vonder Harr 1976) and 4.0\( \text{PW} \) (Masuda 1988). Treberth argues however that there was no scope for further changes in the results since the values of inferred ocean heat transport would lie outside the error bars of direct ocean measurements. The convergence of ocean heat transport values were now calculated to an accuracy of \( \pm 0.3\text{PW} \). Figure 7 from TCO1 has been arguably, the default figure reference on meridional heat transport for the last decade (shown in Fig. 2).

From a more direct approach to measuring ocean heat transport, numerous estimates were made at specific basin sections and latitudes to recreate a 3 dimensional picture of the ocean’s flow pattern. This was the goal of the World Ocean Circulation Experiment (WOCE) with much of its observations taking place between 1990 – 1998. MacDonald and Wunsch (1996) and Wunsch (1998) attempted a synthesis of selected hydrographic sections that covered all major ocean basins to produce consistent meridional heat transport using a global inverse analysis approach. The difficulties arising from this method again originated from sparse spatial and temporal sampling (see section 3.3.2). Koltermann et al. (1999) found evidence of decadal oscillations in meridional ocean heat transport by almost 1\( \text{PW} \) at 36 \( ^\circ\text{N} \) and smaller variations at 24 \( ^\circ\text{N} \) and 45 \( ^\circ\text{N} \) for the North Atlantic basin. However Sato and Rossby (2000) found that aliasing could occur due to inadequate temporal and spatial sampling of seasonal changes. They showed that once the annual cycle is properly taken into account, then interannual variations are within 0.1\( \text{PW} \). The largest source of errors for measurements in the Atlantic basin is due to the variability in the Bering Strait and other regions (MacDonald 1998; Ganachaud and Wunsch 2000).

Estimates of ocean heat transport at 26 \( ^\circ\text{N} \) for the Atlantic region, over the period 1993-2004 (Wunsch and Heimbach, 2006) found no statistically significant trends. Changes in ocean heat transport were found to be coupled stronger to variations in mass transport rather than for potential temperature.

### 3.2.2 Modelling Ocean Heat Transport

With the advent of the new computer now available to the general public, researchers were now starting to experiment in simulations of the climate system (Bryan and Cox 1968; Manabe and Bryan 1969; Sellers 1969) with surprising accuracy to observed measurements, given the
coarse radiation of the model’s grid spacing (up to 5 degrees). It were these early numerical studies, that hinted to man-made induced climate change through our industrial activities. This was further expanded on by Manabe and Wetherald (1979) who conducted one of the first numerical studies that looked at the distribution of climate change related to increasing the CO$_2$ concentration in the atmosphere. Much of the common CO$_2$ induced features of climate change that are seen across the present day IPCC class models were already seen in this early study: the increase in lower troposphere moisture, the reduction of the meridional temperature gradient, the increase in poleward latent heat transport (Fig. 3.2.2b), and the increase in precipitation at higher latitudes (Fig. 3.2a).

![Figure 3.1: Vertically integrated poleward transport of a). Dry Static Energy ($c_p T + \Phi$), b). latent energy ($l_v q$), and c). moist static energy ($c_p T + \Phi + l_v q$), for multiples of pre-industrial CO$_2$ levels. Figure taken from Manabe and Wetherald (1978).](image)

Multiple equilibria of climate states in models were observed by Bryan (1986) and Manabe and Stouffer (1988). These early studies focused on different stable states of the THC which may have impacted on sudden climate change events, such as the abrupt transition between the Allerod and Younger Dryas periods (Manabe and Stouffer, 1988). Other multiple equilibria states have been observed by Emanuel (2002), although the latter used a simple box model
Due to the coarseness of ocean models pre 1990, the ability to model eddy induced motion was a common difficulty. Mesoscale eddies stir the ocean in such a way so that the trajectory of water parcels conserves their density (Bryan et al., 1999). The trajectory of water masses within mesoscale eddies is parallel to surfaces of constant density, allowing for lateral spreading but not mixing across the surface. The early representation of mesoscale eddies as a purely diffusive process in the horizontal direction was therefore insufficient. In this respect, ocean models in native isopycnal coordinate systems had clear advantages over models with constant z-coordinate. An attempt by Redi (1982) and Cox (1987) to parameterize the adiabatic mixing of eddies for z coordinate models used a mixing tensor that mixed tracers along constant density rather than constant depth surfaces. As an improvement, Gent and McWilliams (1990) expanded on the Redi-Cox scheme by introducing an advection due to eddies, on top of their diffusive effects (see section 2.1.7).

As computational power grew, models were now able to reproduce mesoscale eddy activity in the ocean (Jayne and Marotzke, 2002; Meijers et al., 2007). Other advancements led to models requiring no flux adjustments from imperfect interpolation between the ocean and
atmosphere grid e.g. HadCM3 (Pope et al. 2000), leading to better representations of our climate. These were used extensively for the analysis of Global warming scenarios to do with such things as vertical ocean heat transport (Gregory, 2000), atmospheric response to global warming (Kusher et al., 2001), and tropical precipitation changes (Chou and Neelin, 2004) but to name a few.

The idea of Bjerknes Compensation in a Climate Model was investigated by Van de Swaluw et al. (2007) using HadCM3. Using data from a pre-industrial control climate they found model evidence of compensation between the ocean and atmospheric heat transports between 50° – 80°N with the highest correlation observed when the ocean leads the atmosphere by 1 year. Their described mechanism suggests that the AMOC is responsible for controlling the compensation, explained from the following: an increase (decrease) in poleward ocean heat transport will decrease (increase) the equator-pole surface temperature gradient, which in turn would weaken (strengthen) the atmospheric energy transport from baroclinic eddies.

The shape and magnitude of the ocean distribution of meridional heat transport have been previously studied using an idealised aqua planet setup (Marshall et al., 2007; Enderton and Marshall, 2009). By individually adding idealised ridges to mimic land masses, it was found that the distribution of ocean heat transport was dependent on the shape of the continents, in particular the open Southern Ocean Drake Passage. Throughout 4 different model setups, the total heat transport remained relatively insensitive to changes on the aqua planet surface, consistent with Stone (1978).

Model architectural uncertainty has recently come under close scrutiny by Megann et al. (2010). The latter authors examined (though only one section of their work) how the ocean heat transport changes when the ocean model was changed from constant depth to constant density vertical coordinate system, finding small changes that were mainly in within observational errors.

3.3 Methods of Calculating Heat Transport

From the past 100 years into research of heat transport there has emerged two distinct methods for calculating this quantity; these have been termed, the direct and indirect methods. The easiest way to distinguish each method is that the direct method always involves the meridional velocity $v$ in its calculations, whilst the indirect method does not. Both have their own advantages and disadvantages over the other and have evolved predominantly due to the satellite era where the first measurements of a global scale could be finally attained.
3.3.1 Indirect Methods

Figure 3.3: Simple Schematic of Latent Heat transport in the Atmosphere. In steady state, net evaporation $E$ in the tropics and net precipitation $P$ in the extra-tropics leads to a horizontal transport of moisture. This moisture transport can be calculated indirectly from the magnitudes of evaporation and/or precipitation, or directly from the product of wind velocity and specific humidity $L = vq$. A flow of ocean freshwater $F$ is returned equatorward to conserve mass transport.

We consider a simplified two box system shown in Fig. 3.3. As an example this schematic will represent the Atmosphere, with the two regions being the Tropics and the Extra-Tropics respectively. In the Tropics there is net evaporation from the surface, leading to a flux of moisture into the atmosphere. Conversely, the Extra-Tropics is net precipitative, with a flux of moisture into the surface. To satisfy this situation there must be a horizontal transport of moisture from the Tropics towards the Extra-Tropics. Thus the difference between vertical fluxes of moisture ‘indirectly’ calculates a horizontal moisture transport. This example describes the transport of latent heat. The earliest works of ocean heat transport used this method by calculating the changes in the heat storage of the ocean (e.g. Budyko, 1963), with differential heating and cooling requiring, at equilibrium, a transport of heat. An equatorward freshwater flux $F$ is carried out by the ocean to conserve mass.

The advantages of this method have blossomed with the advent of the satellite era. In particular radiation measurements of incoming and outgoing radiation at the TOA and surface used this same method for calculating the heat components of the ocean and atmosphere to a very high accuracy (Trenberth and Caron, 2001; Fasullo and Trenberth, 2008b), drastically changing the magnitudes of the atmospheric heat transport at mid-latitudes that were previously computed using the direct method (see below for details). All components of heat transport are taken into account for the given medium (atmosphere or ocean) giving a ‘com-
plete’ picture that takes into account any missing components that the direct method may miss e.g. diffusive effects. Paradoxically this latter holistic feature for the indirect method is its own disadvantage, not allowing for the various decompositions of heat transport allowed by the next described method.

### 3.3.2 Direct Methods

Again using our two box system, the direct method of calculating poleward latent heat transport would be from the product of meridional wind and specific humidity at the boundary between the two regions. The direct method takes into account the full depth profile of the medium, having the ability to assign specific spatial locations to unique heat transport mechanisms. Within the atmosphere, weather balloons and aircraft based instruments sampled the atmosphere’s properties, giving both limited vertical and horizontal profiles. Similarly in the ocean, sinking measuring instruments such as XBTs were deployed from ships to sample the ocean’s properties.

From ocean vertical profiles of temperature and salinity, one could use the Thermal Wind equations (Eq. 1.18a and 1.18b) to compute the vertical shear of the current. This approach assumes a *level of no motion*, sometimes called a *reference surface* at which ocean current motion is zero, approximately at 2000m (though this value can be highly variable). From this depth, relative currents can be integrated up to the surface and down to the basin floor to give a complete vertical velocity profile (assuming that there were enough measurements of temperature in the first place!).

The advantages of this method are best illustrated for a model with many vertical layers. Using different layers can result in the decomposition of heat transport into various components, e.g. gyre and overturning. This has the added benefit of re-enforcing known theory on ocean circulation mechanisms that are consistent with observations of heat transport.

The disadvantage of this method, particularly in the past was the lack of spatial coverage to measure quantities such as horizontal wind velocity and potential temperature. These were compounded by the poor temporal resolution in measurements. The methods for measurements included numerous errors from instrumental variability and the accuracy of depth profiling - balloons of XBTs never travelled perfectly vertical, but at an angle due to winds or internal ocean currents.

In the past 20 years, there has been a real attempt to improve the global coverage of measuring instruments globally. These included the WOCE and ARGO project (see Chapter
3.3.3 Meridional Heat Transport for the Ocean and Atmosphere

As mentioned in section 1.2 the net radiation perpendicular to the TOA and surface can be a useful quantity in calculating heat transport. The total meridional heat transport carried by the ocean and atmosphere is computed indirectly from the integral of $R_{net}$ in Eq. 1.3 such that

$$H_t = \int_{-\pi/2}^{\phi} R_{TOA}(\phi)2\pi R^2 \cos(\phi) \, d\phi$$

(3.1)

where $\phi$ is latitude. In practice the net radiation at the TOA (in W$m^{-2}$) is adjusted so that the integral over all latitudes results in $H_t$ equalling zero at the poles (Trenberth and Caron 2001). This is accomplished by subtracting the spatial mean value of $R_{net}$ from the entire spatial distribution. For CHIME and HadCM3, these heat storage correction terms account are on the order of $+0.23Wm^{-2}$ and $-0.13Wm^{-2}$ respectively for their control runs scenarios (40 year time mean), listed in section 2.5.1.

Conversely, the ocean component is computed using a similar method by

$$H_o = \int_{-\pi/2}^{\phi} R_{surf}(\phi)2\pi R^2 A(\phi) \cos(\phi) \, d\phi$$

(3.2)

where $A(\phi)$ represents a land mask so that only regions over the ocean are taken into account. The atmospheric component easily calculated as the residual between the two,

$$H_a = H_{tot} - H_o$$

(3.3)

TC01 used the reverse of this method; by using radiative data from the ERBE satellite to compute the net fluxes at the TOA and surface and data from the NCEP-reanalysis and ECMWF, they directly compute the atmospheric and total heat transport, and attained a residual ocean component. A plot of all three components is shown in Fig. 2 with the ocean component computed from NCEP based surface fluxes. Given the differences in land mass configuration and ocean surface between the different poles, the total heat transport remains surprisingly anti-symmetric about the equator with a maximum of about 6PW near 40 °. Much of the poleward heat transport is carried by the atmosphere, particularly at mid-latitudes where it carries 90% of the total heat transport. Peak values of about 5PW at 43 °N greatly exceed previous estimates made by Oort and Vonder Harr (1976) and Masuda (1988) of 3.1 PW and 4.0 PW respectively. Overall the ocean plays a smaller role in poleward heat transport,
reaching a maximum of about 2 PW near 15°N. Equatorwards of this latitude and for the northern hemisphere only, the ocean is actually the dominant career of heat. The significant drop in ocean heat transport near 40°N corresponds well with the latitude of warm western boundary currents (Gulf Stream and Kurushio) turning eastwards, decreasing the meridional component of its heat transport. Similarly in the Southern Hemisphere, the almost negligible heat transport just south of 40 °S is due to the equatorward Ekman transport of surface waters forced by the strong Westerlies within the region.

### 3.4 Types of Heat Transport

The rate of change for the total energy of a large volume can be written as

\[
\rho \frac{D}{Dt} \left( E + \frac{p}{\rho} + \Phi + \frac{1}{2} \bar{u}^2 \right) + \nabla \cdot \left( \bar{F}_{\text{rad}} - k \nabla T - \mu \nabla \left( \frac{1}{2} \bar{u}^2 \right) \right) = Q_H + \frac{\partial p}{\partial t} \quad (3.4)
\]

where we have used the Material Derivative D/Dt defined in Appendix A. The terms on the l.h.s of eq. 3.4 within the material derivative are defined in units per mass as: the internal energy, energy due to compression/expansion of parcel, gravitational potential energy, and kinetic energy respectively. Terms inside \( \nabla \) are defined as the radiative flux, the flux by diffusion of heat, and the flux by diffusion of kinetic energy; \( Q_H \) is the overall heating rate (defined from Gill, 1982, eq. 4.8.2).

The term \( \partial p/\partial t \) is often relatively small (unless examining acoustic waves), along with viscous and diffusive effects except on smaller scales, and therefore can be ignored. In situations where radiative heating and latent heat release can also be ignored, Eq. 3.4 approximates to

\[
\frac{D}{Dt} \left( E + \frac{p}{\rho} + \Phi + \frac{1}{2} \bar{u}^2 \right) = 0 \quad (3.5)
\]

This is known as *Bernoulli’s Equation*.

The quantity \( E + p/\rho \) in eq. 3.5 is often referred to as the *enthalpy* per unit mass. For an ideal gas, this simplifies to

\[
E + \frac{p}{\rho} = c_p T \quad (3.6)
\]

with this approximation used for applications in the atmosphere (Gill, 1982). The addition of the geopotential energy to enthalpy such as in

\[
E + \frac{p}{\rho} + \Phi \approx c_p T + gz = S_d \quad (3.7)
\]

is sometimes called the *Dry Static Energy* per unit mass.
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The Bernoulli Equation (3.5) can be modified to include the effects of latent heat release by approximating $Q_H$ in eq. 3.4 as

$$Q_H \approx -\rho_l \frac{Dq}{Dt}$$  \hspace{1cm} (3.8)

resulting in the modified version,

$$D \frac{Dt}{(E + \frac{p}{\rho} + \Phi + l_v q + \frac{1}{2} u^2)} = 0$$  \hspace{1cm} (3.9)

The quantity

$$E + \frac{p}{\rho} + \Phi + l_v q \approx c_p T + gz + l_v q = S_m$$  \hspace{1cm} (3.10)

is sometimes referred to as the *Moist Static Energy* per unit mass.

Within the atmosphere, the meridional heat transport across a latitudinal circle can be written as

$$H_a = \int \int \rho v (c_p T + \Phi + L) dx dz$$  \hspace{1cm} (3.11)

$$= \int \int \rho v (S_d + L) dx dz$$  \hspace{1cm} (3.12)

$$= \int \int \rho v S_m dx dz$$  \hspace{1cm} (3.13)

where $\Phi = gz$, $L = l_v q$, and all other symbols have their usual meanings.

![Figure 3.4: Total Atmospheric Heat transport (black) decomposed into its Dry Static Energy (blue) and Latent heat (red) transport components. Taken from observations based on NCEP re-analysis data. Figure taken from Pierrehumbert (2002).](image)

The latitudinal distribution of atmospheric heat transport is shown in Fig. 3.4 along with its decomposition into $S_d$ and $L$. In the Tropics, warm air is advected equatorwards in the lower
branch of the Hadley Cell. Since air temperature decreases with height, the net transport of sensible heat $c_p T$ is equatorward. This may seem counter-intuitive however we must also take into account the second term in $S_d$, $gz$. As air rises, this increases its gravitational potential energy such that the change in $S_d$ with height, i.e. $dS_d/dz = c_p dT/dz + g > 0$. Thus the net transport of dry static energy is poleward. Using similar arguments to that of sensible heat, specific humidity decreases with height, resulting in a net equatorward transport of latent heat. It is this partial compensation between poleward moving $S_d$ and equatorward $L$ that inhibits the atmosphere’s ability to transport heat. The atmosphere nevertheless still transports a small amount of heat polewards, though in this region, the majority is transported by the ocean between $0^\circ - 15^\circ N$ (Trenberth and Caron, 2001).

At mid-latitudes the importance of the time mean circulation diminishes with the majority of advection carried by atmospheric eddies. Polewards and equatorwards motion happens at the same altitudes so the vertical gradients associated with these terms are of secondary importance. Atmospheric heat transport is now proportional to $\bar{v}' T'$ and $\bar{v}' q'$. Since atmospheric eddies are to leading order, in geostrophic balance,

$$v' = \frac{1}{f} \frac{d\Phi}{dx} \quad (3.14)$$

where $f$ is the Coriolis Parameter, and $x$ is in the east-west direction, across a complete latitudinal circle, there is no net transport of gravitational potential energy, $v' \Phi' \propto \Phi' d\Phi/dx \approx 0$.

Within atmospheric eddies there is an almost equal contribution of sensible and latent latent being transported poleward, with warm moist air moving polewards, and cold dry air pushing equatorwards.

Further into the extra-tropics, sensible heat transport dominates that of latent heat as the air temperature decreases, reducing the atmosphere’s ability to hold moisture and leading to precipitation.

Within the ocean, the heat transport equation looks very similar to that of Eq. 3.5 but with no latent heat transport term (an atmospheric process). An extensive mathematical work on the energy transport of the ocean is given by Warren (1999) with the use of a similar equation. Warren expands the internal energy component into derivatives for temperature, salinity and pressure, with a cancellation of the geopotential and pressure terms. From the neglect of the kinetic energy term due to its small magnitude, Warren proves that to within
\( \approx 3\% \) the transport of energy in the ocean can be given by the integral

\[
H_o = \int \int \rho c_o \theta dxdz
\]

where \( c_o \) is the specific heat capacity of seawater, and \( \theta \) is potential temperature.

### 3.4.1 Components of Heat Transport

We define the time mean value of a quantity ‘\( X \)’ using an overbar, and the deviations from this mean as primes. Using the same arguments for meridional velocity \( v \), we derive

\[
X = \bar{X} + X'
\]

\[
v = \bar{v} + v'
\]

The time-mean product of these two variables is simply

\[
\bar{vX} = \bar{vX} + \bar{v}X' + v'X
\]

The second and third terms in Eq. 3.18 vanish since the summation of \( \bar{v}' \) and \( \bar{X}' \) is zero. The remaining terms describe motions belonging to stationary and transient motions (Peixoto and Oort, 1992).

We further redefine the time mean of a quantity by its zonal mean value, represented by \( \mathcal{X} \), and the deviations from this spatial mean using an asterix \( \ast \), leading to

\[
\bar{X} = \mathcal{X} + \bar{X}'
\]

\[
\bar{v} = \mathcal{v} + \bar{v}'
\]

The time mean stationary motion can therefore be decomposed into

\[
[\bar{vX}] = [\bar{v}]\mathcal{X} + [\bar{v}']\mathcal{X}'
\]

where the two terms describe the mean meridional circulation and the stationary eddy components respectively.

Combining the results of Eq. 3.19 and Eq. 3.22 we simply get,

\[
[\bar{vX}] = [\bar{v}]\mathcal{X} + [\bar{v}']\mathcal{X} + [\bar{v}']\mathcal{X}^\ast
\]
where reiterating from above, each terms in Eq. 3.23 represents the stationary mean meridional circulation component, stationary eddy component, and the transient eddy component respectively.

In the atmosphere, the time mean stationary meridional circulation is analogous to the Hadley Circulation Cells in the Tropics that keeps its structural integrity throughout the seasons. The time mean stationary eddy component correlates both temporal changes in velocity and the quantity $X$ and is associated with stationary atmospheric waves found at mid-latitude that are forced by orographic features. The transient eddy component correlates temporal changes in velocity and $X$, representing mid-latitude baroclinic storm activities that are constantly developing and evolving.

Similarly in the ocean, the time mean meridional component is analogous to the single cell Atlantic meridional overturning cell. The stationary eddy term describes basin sized lateral gyre motions driven by the winds. Analogous to the baroclinic storms of the atmosphere, the transient eddy component in the ocean describes ocean eddy motion that are typically below 100km across.

The next two sections put into practice our theory and calculations of heat transport for three different climate models. The first part deals with a simple Two-Box Model to show the simple physics of calculating heat transport and its relative sensitivities to varying certain parameters in the model. The second part deals with how heat transport changes with different ocean model structures and how that ultimately feedbacks into the whole system.

### 3.5 Heat Transport in a Two-Box Model

#### 3.5.1 Introduction

Simplified climate box models will always lack the ability to model small scale processes. To save time (both in code development and computational run time) researchers parameterise these processes through simplified equations. One of the most common parameterisations is from equation (2.28) stating that the atmospheric circulation strength is proportional to the meridional temperature gradient multiplied by a constant $K_A$. In EPcm (refer to section 2.4) this constant is tuned to $100/15SvK^{-1}$ - a value that gives a fairly realistic atmospheric mass transport to present day climate. However the fact that certain constants are arbitrary can be seen by some as one of the many Achilles Heels of climate models, subjectively tweaking the control knobs to obtain an idealised climate that may or may not be realistic. On the
other hand there is no substantial evidence as to whether increasing resolution to resolve all physical processes may yield a more realistic climate model. A very coarse climate model with superior parameterisations could in theory thus be a better simulation of our climate than one with high resolution.

Although the views of arbitrary constants may impede the progress of climate models it surprisingly adds another dimension in analysing how sensitive our climate can be to slightly perturbing a few ‘constants’.

In the next sections we will perturb two main parameters to this model:

1. the circulation strength parameter $K_A$
2. the emissivity parameter for water vapour $\gamma$

These two were chosen as were deemed to have the largest impacts on the system. A third parameter (the emissivity parameter for $CO_2$ $\alpha$) does exist however this is not included in the analysis since its variation is simply equivalent to changing the $CO_2$ concentration.

The colours distinguish a certain multiplication of the original parameter value used: blue (0.1), cyan (0.2), green (0.5), black (1.0), yellow (2.0), red (5.0), and magenta (10.0). For brevity we shall analyse the parameters of temperature and specific humidity gradient, atmospheric circulation strength, ocean, atmospheric and total heat transport (including the ratio), and atmospheric latent heat transport. Although results are shown for a fixed $CO_2$ value of 280 ppm (stars), a second set of values are shown that represent results from doubling of $CO_2$ concentration (squares). The inclusion of these results, though not discussed in this section, are included to stop the duplication of figures shown in the later chapters of this thesis. A solid blue line joins up all stars (control climate) for an easier graphical trend representation; a solid red line joins up all squares (double $CO_2$ climate).

A graphical representation of all scenarios described under varying parameter value and $CO_2$ concentration is shown below in Table. 3.1.

An overview of the EPcm model is given in Section 2.4.

3.5.2 Heat Transport Sensitivity to Changing the Circulation Strength Parameter $K_A$

By varying $K_A$ we change the mass circulation strength of both the ocean and atmosphere that are responsible for advecting heat from low to high latitudes.
Global mean surface temperatures (Fig. 3.5a) are warmer (colder) for high (low) values of $K_A$, ranging between $8^\circ C$ ($0.1 \times K_A$) to $22^\circ C$ ($10.0 \times K_A$). Since the circulation strength parameter directly affects the temperature gradient, it is unsurprising that this quantity is indirectly proportional to $K_A$. Specific humidity gradient (Fig. 3.5b) follows a similar pattern to temperature with progressively smaller values of $\Delta T$ and $\Delta q$ needed to raise the global mean surface temperature by a few tenths of a degree.

Although atmospheric mass transport increases with $K_A$ there seems to be two regions where $\Psi_a$ is very sensitive to global mean surface temperatures and one where it is relatively insensitive (Fig. 3.5d). The latter region dominates for colder climates ($K_A$ multiples of 0.5, 0.2 and 0.1) with a decrease in global mean temperature of 9K accounting for a 2/3 reduction in $\Psi_a$. For the warmer regions where $\Psi_A$ is highly sensitive to global mean surface temperatures ($K_A$ multiples of 2.0, 5.0 and 10.0), an increase of 4K relates to an almost quadrupling of mass transport. The different sensitivities relate to the values of $K_A$ being used; for lower range values, $\Psi_A$ easily modulates global mean surface temperatures. However at large values of $K_A$, much more heat is being transported into extra-tropics. At these high limits this region cannot simply hold onto this heat and thus starts to radiate heat away through evaporative cooling. Thus a greater increase in $\Psi_A$ is needed to increase global mean surface temperatures due to this ‘leakage’ effect.

By taking the product of $\Psi_a$ and $\Delta q$ we find that latent heat transport increases (Fig. 3.5c) with global mean surface temperatures, the largest increases of $L(K_A)$ happening between the lowest ($K_A = 0.1, 0.2$) and highest ($K_A = 5.0, 10.0$) values of the circulation strength parameter.

For both the ocean and atmosphere, small (large) heat transports are associated with lower (higher) temperatures that are linked to low (high) multiples of $K_A$ (Fig. 3.6a,b). For the atmosphere and total heat transport there is a positive trend with $T_s$. Surprisingly the oceanic heat transport does not fit this increasing trend, with the distribution reversing between the 0.5 and 1.0 multiple of $K_A$. The trend returns positive between 2.0 and 5.0 multiple. As ocean heat transport is proportional to both the surface temperature gradient (Fig. 3.5a) and the oceanic overturning strength (a tenth the strength for the atmosphere, Fig. 3.5d), the decreases in $H_o$ are linked to multiples of $K_A$ whereby the decrease in $\Delta T_s$ is larger than the increases in $\Psi_O$. The ratio of $H_o/H_a$ has a maximum value for a half multiple of $K_A$, with the distribution decreasing either side of this value, mirroring the non-linear relationship of $H_o$ with $K_A$. 

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Figure 3.5: The sensitivity of certain variables under different multiples of the default circulation strength parameter $K_A$: 0.1 (blue), 0.2 (cyan), 0.5 (green), 1.0 (black), 2.0 (yellow), 5.0 (red), and 10.0 (magenta). Star symbols represent scenarios with prescribed control scenario levels of atmospheric CO$_2$, whilst square symbols represent scenarios with a doubling of CO$_2$. Blue (red) lines pass through all stars (squares).

Figure 3.6: Same as in Fig. 3.5 but for heat transport variables.

3.5.3 Perturbing the $H_2O$ Emissivity Parameter $\gamma$

Emissivity is a property that describes how well an object emits energy through radiation. It is defined as the ratio of energy emitted by the material to that of a black body at the same temperature. An emissivity value of 1 would be a perfect black body; the majority of objects have an emissivity $< 1$. Relating this property to the atmosphere, the main controllers
of emissivity when related to climate change (for EPcm) are CO\textsubscript{2} concentration and water vapour feedback, with increasing concentrations to both these variables leading to a reduction in emissivity. The reduction in atmospheric emissivity results in the trapping of heat within the atmosphere, leading to heating.

Water vapour is by far the most absorbent greenhouse gas of emitted longwave radiation and therefore contributes a large part to the emissivity of the atmosphere. The default value of water vapour emissivity is $\gamma = 1.25$. This value was picked such that the feedback number within a $2 \times CO\textsubscript{2}$ atmosphere relative to a control climate was 0.4, similar to other model water vapour feedbacks (Held and Soden, 2000).

The global mean surface temperature range lies between $-8^\circ C$ ($0.1 \times \gamma$) to $22^\circ C$ ($10.0 \times \gamma$), making the parameter $\gamma$ the primary controller of $T_s$ for colder climates relative to the previous two perturbation scenarios of $K_A$ and $\alpha$.

Just as in perturbations for $\alpha$, the specific humidity gradient $\Delta q$ (Fig. 3.7b) increases with $T_s$ for similar reasons discussed above. What is surprising about both atmospheric mass transport $\Psi_a$ and the temperature gradient $\Delta T_s$ (Fig. 3.7a,d) are that both now increase with $T_s$ up to a certain point, where it then decreases with increasing $T_s$; this turning point value occurs at $0.5 \times \gamma$ (green stars). It is unclear as to the reasons of why the relationship between $\Delta T$ and $T_s$ is parabolic, however it does suggest that the effects of polar amplification are only visual for a certain range of $\gamma$. For multiples larger than 0.5, Polar Amplification is evident with increasing $T_S$. For low multiples (0.1, 0.2 and 0.5) of $\alpha$ the opposite effect happens with an increase in the temperature gradient with increasing $T_s$.

Latent heat transport increases with $T_s$, however due to the parabolic nature of atmospheric mass transport, $L$ starts to decrease with $T_s$ for multiples of $\gamma$ larger and equal to 1.0 (Fig. 3.7c). Repeated calculations of latent heat transport (and the other variables) were made with smaller incremental increases to $\gamma$ (not shown). To better understand this peak, we find that by using this finer resolution, the turning point multiple value for $L$ happens at $0.6 \times \gamma$, with latent heat transport starting to decrease (increase) with CO\textsubscript{2} with a multiple greater (less than) this value. It is worth noting that there is very little change between quantities within these high multiple ranges, inferring that the effects of water vapour feedback become increasingly saturated as global mean surface temperatures are at the highest range of $22^\circ C$. This would be consistent with a saturated water vapour climate.

Due to the quadratic distribution of $\Psi_a$, ocean, atmospheric and total heat transport exhibit similar distributions with peak maxima occurring for $0.5 \times \gamma$ (Fig. 5.11a,b,c). The
ratio of ocean to atmospheric heat transport mostly increases with $T_s$ (Fig. 5.11d) apart from a transitionary region between multiples 0.5 and 1.0 where the trend reverses, this time due to the decrease in $H_o$ being greater than the decrease in $H_a$.

Figure 3.7: Same as in Fig. 3.5 but for varying $\gamma$ dependence.

Figure 3.8: Same as in Fig. 3.6 but for varying $\gamma$ dependence.
3.6 Comparisons of Meridional Heat Transport for CHIME and HadCM3 control scenarios

3.6.1 HadCM3 and CHIME model setup

The model setup for both CHIME and HadCM3 are explained in Chapter 2. For an extensive description see Megann et al. (2010).

3.6.2 Heat Transport

We infer the total heat transport $H_{tot}$ (ocean and atmosphere) from the divergence required to balance out the net radiation at the top of the atmosphere (TOA), using Eq. 3.1. We note that $R_{TOA}$ is adjusted to remove any heat storage terms by subtracting the spatial mean net radiation at the TOA, calculated at $-0.13\, Wm^{-2}$ and $+0.23\, Wm^{-2}$ for HadCM3 and CHIME respectively. It is worth noting that if $R_{TOA}$ does not change, $H_t$ will remain invariant whereby changes in either the ocean or atmosphere will be compensated by opposite changes within the other - this is the basis of the Bjerknes Compensation (Bjerknes 1964).

For brevity, any described differences in a quantity ‘X’ between each model will hereafter be such that $\Delta X = X_{chi} - X_{had}$, where the subscripts represent ‘X’ values for each model. Note that ‘X’ is different from the one used in section 3.4.1.

![Figure 3.9: Meridional heat transport curves for HadCM3 (black) and CHIME (grey). Curves for the total, atmosphere and ocean are given by the solid, dashed and dotted lines respectively (A). Atmospheric heat transport is further decomposed into its dry static energy and latent heat components represented by the dot-dashed and solid square lines respectively (B).](image)
Chapter 3

Figure 3.10: Differences in the meridional heat transport (CHIME - HadCM3) from curves shown in Fig. 3.9. Inter-model differences show a partial compensation between the ocean and the atmospheric components (A). The largest changes are seen for the DSE and L components of the atmosphere at 7.5°N due to a northward shift in the ITCZ for CHIME (B).

Fig. 3.9A plots the distribution of $H_t$ for HadCM3 (black, solid), CHIME (grey, solid), with Fig. 3.10A showing the difference $\Delta H_t$ (solid). Both model distributions of $H_t$ are similar, with differences of the order of $\Delta H_t = 0.2PW$ being dominant in the Tropics. Though this results in CHIME having a slightly larger total heat transport, this is of secondary importance as compared to changes in the ocean $\Delta H_o$ (Fig. 3.10A, dotted) and atmosphere $\Delta H_a$ (Fig. 3.10A, dashed). We can therefore view both models as having to first order, the same total heat transport. The main importance of this feature is that the total heat transport constrains any changes such that differences within the ocean and atmospheric heat transports must balance out for the net effect to be approximately zero. Although the principles of Bjerknes Compensation originally refer to changes in heat transport for one model evolving within its own environment, the applications are the same for our two model comparison of the same control climate.

For HadCM3, ocean heat transport is calculated using the same method as in Eq. 3.2, whereby $R_{surf}$ is the net input of radiation at the ocean surface, with zonal mean values weighted by the fraction of ocean to land surfaces at constant latitude. For CHIME, $H_o$ is calculated directly from the summation of meridional velocity $v$ and potential temperature $T$ across a zonal cross-section. This method was used for CHIME since the non-conservation of heat within its ocean component led to spurious errors when using the method used by
As the CHIME ocean grid is bipolar north of 55°N the identification of ocean grid points at certain latitudes above this point becomes increasingly difficult due to the curved nature of the grid. Calculations of \( H_o \) from the \( v \) and \( T \) thus lead to a ‘saw-tooth’ distribution of heat transport. To minimise this ‘noise’ associated with the curved surface we employed a Savitzky-Golay algorithm (Barak, 1995) to smooth out the calculated heat transport distribution at high latitudes. Since values of \( H_o \) are particularly small for this region compared to within the Tropics, they are thus of secondary importance within this paper. Both curves of \( H_o \) (dotted) are shown in Fig. 3.9A, and are in close agreement to that calculated by M2010 (their Fig. 5). Atmospheric heat transport \( H_a \) (Fig. 3.9A, dashed) was calculated as a residual between \( H_t \) and \( H_o \). The direct method for calculating \( H_a \) (similar to that of \( H_o \) in HadCM3) was not chosen due to the poor temporal resolution of the data sets. The monthly mean time steps would be too coarse to resolve mid-latitude transient weather patterns i.e. storms, thus any direct calculations of \( H_a \) would underestimate these regions; daily data would therefore be needed.

From Fig. 3.10A there is indication of compensation between the ocean and atmospheric transports. Whereas HadCM3 has a larger ocean heat transport magnitude, CHIME’s atmospheric heat transport also increases to compensate.

The distribution between \( \Delta H_o \), \( \Delta H_a \), and \( \Delta H_t \), although similar, are slightly larger than observed model changes due to anthropogenic forcing. These document maximum changes in total heat transport by 0.05\( PW \), and less than 0.2\( PW \) for changes in the ocean and atmospheric heat transport (see Hwang et al. 2011, their Fig. 3, which includes the corrected figures to Held and Soden, 2006). Changes in the choice of vertical coordinate system (and corresponding mixing scheme) for the ocean are thus larger than those resulting from anthropogenic forcing.

The atmospheric energy transport is defined by Eq. 5.6.

We compute latent heat transport indirectly from the differences in evaporation minus precipitation at the ocean surface. Analogous to the net radiation at the TOA for computing \( H_t \), the divergence (evaporation) of moisture to an area of convergence (precipitation) must lead to a horizontal transport of freshwater in the atmosphere. Equation (3.2) is therefore used, with \( R_{surf} \) being replaced by the zonal mean of the \( E - P \) spatial distribution. The DSE component is calculated as a residual between \( H_a \) and \( L \). For both hemispheres, the transport of latent heat within the Tropics is equatorward.

Fig. 5.2B shows the latitudinal distribution of \( L \) for HadCM3 (black line, square), CHIME
(grey line, square), with the difference $\Delta L$ shown in Fig. 3.10B. Both curves for each model have consistently similar magnitudes, except at the deep NH Tropics ($0-20^\circ N$) where HadCM3 is larger by $\approx 1.2PW$. This difference is nearly the same order of magnitude as that of the ocean heat transport and thus warrants further investigation. The magnitude and location of the maximum value of latent heat transport in the NH Tropics is $-1.3PW$ at $7.5^\circ N$ and $-0.6PW$ at $10^\circ N$ for HadCM3 and CHIME respectively, suggesting a northward shift of the ITCZ region by $2.5^\circ$, as noted in M2010.

The changes to the transport of DSE is most obvious at $7.5^\circ N$, with its size being of similar magnitude to the changes seen in $L$. Since $|\Delta H_a| << |\Delta DSE|$ and $|\Delta L|$, the atmospheric heat transport effectively constrains the changes to both these latter components. This however is not like the classical Bjerknes Compensation for the ocean and atmosphere whereby each term changes by an equal but opposite amount. For DSE and $L$, both quantities decrease by similar amounts. This is perfectly plausible since at these low latitudes, the transport of DSE and $L$ are in the opposite direction to each other, leading to a small net transport of heat.

The interpretation of Fig. 3.10 is that although there are compensating changes within the ocean and atmospheric heat transport, the largest compensation is seen within the atmospheric internal decomposition. It is thus surprising to find that switching to a different vertical coordinate system (an ocean property) could have such large feedbacks within the atmosphere. For the rest of this paper we therefore aim to explain this reduction in latent heat and DSE transport and how we can relate this back to the ocean coordinate system in use.

Since the dominating structure of meridional heat transport within this region is the Tropical Hadley Cell we next analyse its properties to determine where this decrease originates from.

### 3.6.3 The Northern Hemisphere Hadley Cell

A summary of the Hadley Circulation is given in Chapter 1, section 1.3.1.

The atmospheric mass stream function is defined by

$$\Psi_a = \int \int vdx \frac{dp}{g}$$

(3.24)

where the integrals are over all longitude and pressure. We construct $\Psi_a$ using meridional velocity wind profiles interpolated onto constant pressure levels at 10hPa divisions. By applying mass conservation around a latitudinal circle, we integrate from the TOA ($p=0$) to a pressure $p$. Fig. 3.11A and 3.11B shows $\Psi_a$ for CHIME and HadCM3 respectively, highlighting the dominant circulation patterns of the HCs in the Tropics, and the weaker mid-latitude
Figure 3.11: Meridional atmospheric mass stream function $\Psi_a$ in $kgs^{-1}$ for CHIME, highlighting the dominant Hadley Cell within the Tropics, and the weaker Ferrel and Polar Cells at mid and high latitudes respectively (A). The difference in $\Psi_a$ between CHIME - HadCM3 (B). The ‘anti’ Hadley Circulation signifies CHIME having a weaker circulation strength to that of HadCM3 by almost 50%.

Ferrel Cells and the high latitude Polar Cells. At the centre of the NH HC the maximum value of mass transport is $\Psi_{a}^{max} = 6.6 \times 10^{10} kgs^{-1}$ and $\Psi_{a}^{max} = 9.7 \times 10^{10} kgs^{-1}$ for CHIME and HadCM3 respectively.

We model the freshwater transport in the atmosphere $F_a$ as the product of the atmospheric mass stream function $\Psi_a$ and the specific humidity of the lower troposphere $q$ such that

$$F_a = \Psi_a q \quad (3.25)$$

Here $q$ is the fraction of water vapour in the atmosphere with respect to its total mass. Inspection of the latitude height distribution of $q$ (Fig. 3.12A and 3.12B) shows the highest values near the equatorial and tropical regions, consistent with areas of highest surface temperatures having a moisture rich atmosphere. Values of $q$ decrease poleward and upward as the air temperature decreases, resulting from a reduction of atmospheric moisture through precipitation and less upward transport. Fig. 3.12C illustrates the difference between the meridional profile of $q$, highlighting CHIME having larger values at the tropical regions,
especially for the NH. These results are consistent with CHIME having a warmer surface temperature, and increases in $q$ being within the latitudinal region of interest. The largest values of $q$ for CHIME, HadCM3 and the model differences are on the order of 17g/kg, 16g/kg and 1.5g/kg respectively, resulting in a residual increase of 9% for low level values.

As described previously, the atmospheric mass stream function is shown in Fig. 3.11A and 3.11B. The differences in $\Psi_a$ between the two models (Fig. 3.11C) reveals a residual circulation within the NH Tropics, the sign of which is in the opposite direction to the regular Hadley Circulation. This ‘anti’ Hadley Cell has a maximum mass stream function of $-5.0 \times 10^{10} kg s^{-1}$ and is associated with a reduction in the circulation strength of the NH Hadley Cell of CHIME relative to HadCM3 by approximately 52%. The latitudinal expanse of the residual circulation is within the same location as the residual increase in $q$. Although there is a smaller ‘anti’ Hadley circulation within the region of the southern hemisphere HC, its magnitude is small compared to that of the actual circulation itself, and thus represents minimal changes to $\Delta L$.

From comparisons of both $\Psi_a$ and $q$ we therefore find that the decrease in latent heat transport of CHIME within the NH Tropics originates primarily from the reduction in the
atmospheric mass transport (−52%) being larger than the increase in specific humidity (+9%) from warmer SSTs. This dynamically driven reduction in $L$ by $\Psi_a$ is consistent with the reduction in DSE since both processes are proportional to the strength of the atmospheric circulation. The decrease in $\Psi_a$ by almost half also agrees well with equatorward maximum of latent heat transport for the NH, given at $−1.3PW$ and $−0.6PW$ for HadCM3 and CHIME respectively.

A relevant question to ask from this conclusion is by what mechanism would the Hadley Cell Circulation decrease? To answer this we are reminded that the Hadley Cells are thermally driven (see section 1.3.1), meaning the meridional temperature contrast across the Hadley Cell region is an important indicator of the strength of $\Psi_a$ (Gill and Rasmusseen, 1983; Neelin and Held, 1987). The highest SSTs are found near the equatorial region, away from which its magnitude decreases with latitude. There is thus a meridional temperature gradient across all latitudes.
Figure 3.13: The difference in sea surface temperature between CHIME - HadCM3. CHIME has a predominantly warm bias SST compared to HadCM3 as a result of its chosen ocean z-coordinate system and complimentary mixing scheme. Increased differences in SST are localised near mid-latitude western boundary current regions that act to decrease the overall temperature gradient of CHIME across the Tropics.

Figure 3.13A and 3.13B shows the SST anomalies from the mean NOCs climatology for CHIME and HadCM3 respectively. These are shown instead of the raw SST figures as they highlight in more detail the discrepancies each model have with simulating present day climate. Fig. 3.13C illustrates the SST difference between CHIME and HadCM3, showing the ever-present warm bias of CHIME’s SST relative to HadCM3. This warming is non-spatially uniform with CHIME having warmer western boundary currents such as the Gulf Stream (North Atlantic) and the Kuroshio (North Pacific). From inspection, the relative warming in the North Pacific is primarily related to HadCM3 being colder within the region, rather than CHIME being warmer. Near the Gulf Stream, the combination of a warmer CHIME and colder HadCM3 leads to the overall warm bias in that region. The larger differences in SST at mid-latitude as compared to the equator, acts to reduce the meridional temperature gradient across the NH Tropics in CHIME, which we hypothesized leads to the reduction in $\Psi_a$.

Sea surface temperature is therefore the variable that links the ocean model’s architec-
tural properties (vertical coordinate system and mixing schemes) to the atmosphere, with surprisingly large feedbacks.

Hadley Cell dynamics have been well documented (see Walker and Schneider 2006, and references), describing a thermally driven meridional circulation that is proportional to the North-South temperature gradient at the surface. From comparisons of SST maps between both models it was revealed that CHIME’s SSTs had an overly warm bias. This was particularly evident for the mid-latitude Western Boundary Currents of the North Pacific and Atlantic regions. The uneven spatial distribution of SST for CHIME led to a reduced surface temperature gradient across the Tropical Northern Hemisphere that resulted in the reduced atmospheric mass transport of the Hadley Cell.

3.7 The Hydrological Cycle

The Hydrological Cycle deals with the transport of freshwater in the climate system, merging the water processes from land, sea, and air, and covers all states of water, from solid ice, liquid precipitation, and gaseous moisture. The majority of the water carried by the atmosphere originates from the ocean. The exchange at the ocean-atmospheric interface results in distilling (evaporation) or freshening (precipitation) of the ocean surface. The salinity budget of the ocean thus reflects these exchanges and can be used as an indirect proxy to the processes in the atmosphere. In steady state, freshwater transport in the atmosphere is of equal and opposite direction within the ocean to conserve mass across a latitudinal boundary. At mid-latitudes, freshwater transport in the atmosphere accounts for roughly half the total amount of its poleward heat transport. Understanding the freshwater processes in the ocean is thus no trivial a task.

The amount of water in the ocean is approximately $1.4 \times 10^9 km^3$, which is about 100,000 times the water storage capacity of the atmosphere (Schmitt, 2008). A global estimate of the total evaporation is about 13 Sv with the total precipitation over the ocean totalling 12 Sv. The 1 Sv difference is the portion due to continental runoff back into the ocean, and is equivalent to the difference between total land precipitation of 3 Sv minus the total evapotranspiration of 2 Sv. Table 3.14 decomposes the different contributions of the global water cycle into segments belonging to different ocean basins, and components belonging to the major global rivers.

A brief look at the fluxes reveal that the hydrological cycle is dominated by ocean-
atmospheric interactions, with very small contributions from land and glacial melt discharge. These magnitudes should not be taken as absolute values but contain large uncertainties due to the limitations of our present observations (see below).

### 3.7.1 Indirect Estimates of Freshwater Transport

Estimates of surface and atmospheric freshwater fluxes from satellites and near terrestrial sources have been used to measure variables associated with evaporation, precipitation, or atmospheric moisture content (Xie and Arkin, 1997). From the use of data-assimilating atmospheric models there have been numerous problems in assessing their validity: generally the model outputs did not conserve total mass (Trenberth and Guillemot, 1998), and secondly, the lack of atmospheric data over the ocean; the latter leading to spatial biases centred on available observation centres.

Estimates of surface evaporation are computed from empirical relations for variables such as wind velocity, relatively humidity and surface temperature, measured either via satellite, ships, or land stations. With the formulae used not being perfect, estimates of evaporation do suffer from biases especially when integrated over large areas.

Precipitation measurements over the ocean have been particularly difficult. Direct observations from island stations can suffer from small-scale anomalies due to local orographic features, leading to random biases. A more ‘complete’ technique has utilised the use of satellites that use microwave radiometers to penetrate clouds and measure the radiation emitted from rainwater and the scattering caused by cloud ice and snow (Xie and Arkin, 1997). Al-

![Figure 3.14: Components of the water cycle in \( \text{km}^3/\text{yr} \) and in Sv (1Sv = 10^6 m^3/s). Table taken from Schmitt (2008).](image-url)
though much more progressive, this method relies on empirical algorithms that require ‘tuning’ to real world observations. The limitations of satellite exposure to the ocean surface presents its own set of biases when calibrating data (Wijffels, 2001).

The meridional freshwater transport in the atmosphere, or latent heat transport can be calculated by a similar equation to 3.1 whereby global zonal mean values of evaporation minus precipitation are integrated from the South Pole up to a given latitude. Mass conservation is enforced whereby the global mean value of E-P is removed from the spatial distribution such that the integration of latent heat transport remains zero at both poles. Put different, mass conservation requires that net evaporation equals net precipitation.

### 3.7.2 Direct Estimates of Freshwater Transport

Within the atmosphere direct calculation of freshwater transport involve the integration of wind velocity and specific humidity across zonal cross-sections similar to that of eq. 5.6 where $L$ is $l_v q$. Just as for inferred measurements of ocean velocity, atmospheric velocities rely on the thermal wind equation and geostrophic dynamics.

The direct computation of ocean freshwater transport is similar to that of heat, utilising enclosed ocean volumes for which budget equations can be written. It assumes that we can measure the steady-state fields of velocity and salinity, and that geostrophy dominates within the ocean interior, except for the Ekman layer (Wijffels, 2001). Changes in the velocity and salinity field at the seasonal or meso-scale eddy level are ignored as they have been found to be small over most of the ocean (Wijffels, 2001).

For an enclosed hydrographic section, the balance of salt transport in the ocean and atmosphere in steady-state implies that

$$\int \int \rho S \, v \, dx \, dz = T_i^S \quad (3.26)$$

where $\rho$ is in-situ density, $T_i^S$ is the interbasin transport of salt eg. Bering Strait, and all other symbols have their usual meaning.

Similarly the mass conservation across a similar hydrographic section can be written as

$$\int \int \rho v \, dx \, dz + |P - E + R| = T_i^M \quad (3.27)$$

where $T_i^M$ is the interbasin mass transport, and P, E, and R are the precipitation, evaporation and continental runoff respectively.
Since seawater is a dilution of salt and freshwater, the latter term can thus be expressed as the difference between the two mentioned equations such that

\[
\int \int \rho (1 - S) v dxdz + [P - E + R] = T_i^M - T_i^S
\] (3.28)

where the L.H.S of eq. 3.28 is strictly, the ocean freshwater transport.

Additionally we define the section area average value of salinity computed as

\[
\bar{S} = \frac{\int \int S dxdz}{\int \int dxdz}; S' = S - \bar{S}
\] (3.29)

As a simplification we assume that the interbasin salt flux occurs at a known salinity \( S_i \), such that we can express Eq. \( s T_i^S = S_i \times T_i^M \). By combining Eq. 3.26, 3.27 and 3.29 we generate

\[
[P - E + R] = -\frac{T_i^M S'_i - \int \int \rho S' v dxdz}{S}
\] (3.30)

where the terms on the R.H.S of Eq. 3.30 are defined as the ‘leakage’ term associated with the cross-sectional area transport equal to the interbasin exchange, and the correlation of salinity and velocity across the cross-section (Wijffels, 2001). As stated in section 3.3.2, \( v \) is measured from geostrophic and Ekman dynamics from density and wind stress profiles, whilst \( S \) is measured in-situ. The errors associated with Eq. 3.30 have been found by Wijffels (2001) to be on the order of 0.17 Sv outside of the Tropics, and as much as 0.3 Sv within the Tropics linked predominantly to uncertainty within the wind-driven components.

### 3.8 Comparison of the Hydrological Cycle in CHIME and HadCM3 within a Control Climate Scenario

#### 3.8.1 Methodology

The spatial patterns of evaporation (E), precipitation (P), and the differences (E-P) are shown in Fig. 3.15, 3.16, and 3.17 respectively for both the models CHIME and HadCM3. Units are in \( Wm^{-2} \) with positive (negative) values corresponding to fluxes out of (into) the surface.

Evaporation rates peak in the Tropical regions and decrease with latitude, following a similar spatial distribution to surface temperatures. The greatest evaporation rates are found in the open ocean where there is lowest albedo and an abundance of freshwater, in particular, the regions of western boundary currents. Evaporation rates over land are weaker, having a higher surface albedo due predominantly to vegetation and snow cover. Extremely dry
land regions such as the African Sahara Desert have almost negligible evaporation since there are no surface surfaces for moisture to originate from. Both models possess similar spatial distributions of evaporation. Model differences have a very broad spatial distributions, with CHIME having slightly larger evaporation rates by about 7%. The zonal mean distribution of evaporation suggests that many of these model variations occur within the mid-latitude and tropical areas of the models.

Model precipitation rates are greatest in the equatorial belt, characterised by the wet ITCZ region, and at extra-tropical latitudes by storm-tracks, consistent with Hadley Circulation dynamics and poleward transport of moisture. Precipitation rates have been multiplied by \( l_v \) to give units of \( W m^{-2} \). Rates are lowest in the Tropics where evaporation tends to dominate. Only inside the path of ocean western boundary currents, transporting heat in a north-easterly direction for the northern hemisphere, does precipitation have a substantial signal. The strong precipitation signals around these areas may correlate well with the position of the mid-latitude storm tracks that can anchor to the positions of the western boundary currents (Hoskins and Valdes, 1990). This would be consistent with the high evaporation rates within the region aiding in the formation of moisture rich clouds that are blown north-eastwards, eventually precipitating on its way. Just as evaporation, spatial patterns of precipitation between both models contain similar structures. The differences between the models however reveal a striking cross equatorial dipole pattern that describes a residual northward of transport of moisture near the surface. That is to say, that CHIME has a slightly northward shift in the position of its ITCZ region compared with HadCM3 by 1 atmospheric grid point, or 2.5°.

E-P reveals that the Tropics are the net source of moisture to the atmosphere, with the net sinks located at the equatorial ITCZ region and extra-tropical latitudes. We note that the zonal mean integrations for E-P include the land areas which have been masked in the plots. The differences of E-P between models share a high correlation to the spatial differences of precipitation, rather than for evaporation. There is a northward shift of the ITCZ region in CHIME by a one grid atmospheric resolution of 2.5°N, discussed in section 3.6 to being due to a weakened northern hemisphere Hadley Cell that was driven by a weaker SST gradient across the tropical region. If we use location where zonal net evaporation changes to net precipitation as a physical indicator to the tropical boundary (see Chapter 1, section 1.5.4), Fig. 3.17D indicates that for the NH, CHIME has a wider tropical boundary by approximately 2° (centred at 37.2°N) when compared to HadCM3; in the SH the difference is substantially smaller with CHIME increasing its latitude boundary by approximately 0.5° (centred at 40.2°S). The
hemispherical differences in this physical indicator to the boundary of the tropical region is consistent with CHIME having anomalous high SSTs in the NH relative to the SH, that may contribute to this effect.

Figure 3.15: Evaporation rates for A) CHIME, B) HadCM3, C) CHIME - HadCM3, and D) their zonal mean. Data represents a 40 year mean period between years 80 - 119 of a control climate scenario.

Figure 3.16: Precipitation rates for A) CHIME, B) HadCM3, C) CHIME - HadCM3, and D) their zonal mean. Data represents a 40 year mean period between years 80 - 119 of a control climate scenario.

We decompose the zonal mean distribution of E-P in Fig. 3.18 into areas associated with the Atlantic basin, Indo-Pacific basin, Southern Ocean, and a residual land component. Each component has been weighted by the fraction of its area contained at each latitude. Results are shown both for CHIME (dashed) and HadCM3 (solid). Due to the sheer size of the Indo-Pacific region, the global mean correlates the highest to the latter region, with the Atlantic playing
a secondary role in the overall picture. Area weighted evaporation rates for the Atlantic are approximately half that for the Indo-Pacific region, with almost negligible contributions to precipitation in the ITCZ and mid-latitude regions across both models. The Southern Ocean, defined as all areas south of 30°S is virtually identical to the global mean since the region circumvents the entire globe. The land component constitutes a very small fraction of the total net E-P per latitude, not surprising since land surfaces are only a third of the global area.

Comparing both models, there is indication that HadCM3 has higher evaporation rates per latitude in the northern hemisphere Tropics than CHIME, again consistent with a weaker equatorward transport of moisture in CHIME for the region.

To gain a sense of how much freshwater is entering or leaving an ocean basin from simply evaporation and precipitation rates, we sum up the individual latitudinal components for each region shown in Fig. 3.18, with results shown in Table. 3.2. To reiterate, positive (negative) values correspond to net evaporation (precipitation) for the entire region. Values have been rounded to two decimal places.

A significant observation is that although HadCM3 has increased evaporation in the northern hemisphere Tropics, its net precipitation near the ITCZ offsets this feature, leading to an overall reduction in evaporation for a basin average. In comparison, CHIME is almost completely net evaporative even near its ITCZ, leading to an overall increase in net evaporation. This point is expanded and discussed further below.
Both the Atlantic and Indo-Pacific regions are net sources of moisture to the atmosphere. Although the Atlantic covers a smaller area than for the latter region, it has a greater overall net evaporation rate, supporting both model and real world observations of higher surface salinities for the Atlantic region rather than the Indo-Pacific. The Southern Ocean is net precipitative, with moisture converging into the region from all the major basins within the Tropics. Land is also net precipitative, having no substantial internal sources of water. Although the latitudinal distribution of freshwater flux on land is small compared to the other ocean areas, its large latitudinal extent results in an overall precipitation rate that is larger than for the Atlantic or Indo-Pacific basins. Ultimately this land freshwater flux returns back to the oceans via river runoff or subterranean leakage.

Both HadCM3 and CHIME show a high resemblance to computed freshwater fluxes, in particular for the Southern Ocean where differences are less than 2%. This is surprising since HadCM3 has been shown to have a cold SST bias within the region whilst CHIME is overly warm (Megann et al. 2010). It may however mean that SST does not play an overly large role in affecting moisture flux within the region. A striking feature amongst the data is shown in the model differences for the Atlantic and Indo-Pacific regions; although in both models, the two regions are net evaporative, CHIME has a slightly higher evaporation rate for the Atlantic (+0.04 Sv) and a slightly less evaporation rate for the Indo-Pacific region (−0.05 Sv). The very similar magnitudes coupled with their close proximity to each other would suggest that there could be an anomalous transport of freshwater from the Atlantic into the Indo-Pacific basin - this is studied below. Since we have only looked at vertical fluxes of evaporation and precipitation within each region, it can be only the atmosphere that transports this moisture into the region and not by the ocean.

For HadCM3, runoff is computed from the river discharge variable 'outflow' with an additional component from the flux of melt water at the bottom of glaciers, weighted by the ice fractional area. For CHIME, the method for calculating runoff is identical except that there is no variable for continental runoff (without direct ice melt contribution). The available variable used was pmep, representing the ‘water flux into sea water from rivers and surface downward water flux’. To isolate the ‘runoff’ variable we subtracted the components of evaporation and precipitation from ‘pmep’. All CHIME ocean variables are then interpolated onto the atmospheric grid.

The inclusion of continental runoff into our ocean basin freshwater budget calculations are shown in Table. 3.3. We note that for both models, individual data points of runoff
have been multiplied by empirically derived values such that the spatial summation equals that to the total E-P rate over land. That is to say that all land precipitation returns into the oceans. For HadCM3 this correction is extremely small since the discrepancy between calculations of runoff and land E-P rates are only 0.17%. For CHIME however, the discrepancy is 5.8% between the two calculations (1.16 Sv and 1.23 Sv for integrated land E-P and runoff calculations respectively), needing a ‘correction’ factor of 0.0448 to achieve congruence. The large model differences are predominantly due to the curved grid that CHIME’s ocean model is based on; variables of runoff for CHIME are imported from the curved ocean model grid and interpolated onto the squared atmospheric grid. The interpolation is not perfect (personal communication from A. Megann) with coastal areas on the interpolated atmospheric grid being slightly different to the original, resulting in a loss of information. These irregularities are only limited to regions northwards of 55°N where the ocean grid starts to become curvilinear. Given the furthest northward extent of the Atlantic and Indo-Pacific basins is 70°N and that everything above this latitude is grouped under $F_l$, the author feels that to a first order approximation, that this correction is valid for including continental runoff rates into each individual basin.

With the addition of continental runoff into the freshwater budget of each basin, we find in both models, a substantial reduction in the overall net evaporation rates of the Atlantic and Indo-Pacific basins to approximately half and a third of values respectively, to rates without runoff. Both regions receive approximately $-0.6\,Sv$ each, equal to roughly half the total land runoff component. There is a slight increase in net precipitation for the Southern Ocean due to the small land mass of the South America, south of 30°S. The small ‘land’ component shown for values of $F_l$ is not an error, but associated with the combined land and sea areas northward of 70°N.

Even with runoff, the model differences reveal that CHIME’s Atlantic region has an increased evaporation rate by about 0.04Sv. With the Indo-Pacific and the Southern Ocean region being net precipitative by $-0.01Sv$ and $-0.04Sv$ respectively, we cannot say definitely that the majority of evaporative freshwater from the Atlantic is transported into the Southern Ocean; since continental runoff is included there exists a component a substantial component from the Antarctic peninsula that adds freshwater into the region from the south. Additionally the time mean pathways of atmospheric advection near the tropical regions suggest a substantial east-west transport rather than north-south, particularly near the equator. Concentrating solely on E-P without continental runoff, the anomalous differences shown must therefore be
transported by the atmosphere. We hypothesise that there is a small anomalous moisture transport of approximately $0.05\,\text{Sv}$ from the Atlantic into the Indo-Pacific region. Furthermore, we hypothesise that this pathway is westwards through the Panama Isthmus near the equator (Broecker, 1991).
3.8.2 Moisture Transport through the Panama Isthmus

As shown from table 3.2 there is a strong basis to assume that there is an anomalous freshwater transport in CHIME, from the Atlantic into the Indo-Pacific region. The mechanism for this is proposed as follows:

- CHIME model has higher sea surface temperatures relative to HadCM3
- excess heat increases rate of evaporation, adding extra moisture into the atmosphere
- smallest land mass separating the Atlantic from the Indo-Pacific basin is through the Panama Isthmus near the equator
- CHIME’s excess moisture is transported through Panama Isthmus westwards via the Trade Winds
- any eastward transport of moisture from the Pacific back into the Atlantic via the Westerly Winds is blocked by orographic features

The above sequence thus describes a one-way transport of moisture that heavily relies on orographic features and time mean atmospheric pathways; a schematic of this mechanism is
shown in Fig. 3.19. To test our hypothesis we directly compute the zonal atmospheric moisture transport through a latitudinal range between 0°N and 20°N, roughly consistent with the extent of Central America. Longitudinal coordinates were varied with latitude to correspond to the first grid space westwards from continental America, in other words, ‘hugging the coast’ (represented by thick purple line in Fig. 3.19). Zonal mean atmospheric moisture transport $F_u$ was computed from

$$F_u = R \int_0^{p_0} \int_0^{2\pi/9} u q \frac{dp}{g} d\phi$$

(3.31)

where $u$ is zonal wind speed, $d\phi$ is the latitudinal integrated between 0 (equator) and $2\pi/9$ (20°N) equal to approximately $2.2 \times 10^6 m$, and all other symbols have their usual meaning. Our calculations use mean sea level pressure to take into account orographic features and not just the surface reference pressure for each model. Results for $F_u$ were calculated at $-0.43 \pm 0.14(0.03)$ Sv and $-0.39 \pm 0.18(0.07)$ Sv for HadCM3 and CHIME respectively; the first set of error bars denote the standard deviation using 40 years of monthly mean data, whilst the number in brackets is the standard deviation using 40 year annual mean data. The difference between the two ie. CHIME - HadCM3 yielded an anomalous transport of $+0.04 \pm 0.32(0.10)$ Sv. This result shows that instead HadCM3 transport more moisture westwards than CHIME, the complete opposite to our original hypothesis (needed -0.04 to -0.05 Sv). Although the size of the errorbars does open the possibility that CHIME could have a stronger freshwater transport, the author feels that this may be ‘stretching the truth’ using statistics, especially when the standard deviations are approximately one order of magnitude larger than its mean variable.

Different latitudinal ranges were also tried (20°S-20°N, 10°S-20°N, 0°N-10°N), though all resulted with similar conclusions. Due to the small model differences we are trying to calculate, the model internal variability is too large to draw any meaningful conclusion from these results.

Given the results, our original hypothesis was proved invalid. Nevertheless, this mechanism is tested further within an anthropogenically forced climate scenario where $CO_2$ concentrations increase up to 4 times pre-industrial values.

### 3.9 Chapter Summary

This Chapter focused on three main applications of heat and freshwater transport, namely 1). the sensitivity of heat transport in the two box model EPcm to varying certain parameters, 2).
Figure 3.19: Schematic of Anomalous Freshwater transport through the Panama Isthmus from the Atlantic into the Indo-Pacific Ocean. The red arrow corresponds to moisture transport carried by the Easterlies. Blue horizontal arrows represent freshwater transport directed back towards the Atlantic via the Westerlies. The moisture is eventually blocked by orographic features (brown rectangles) of the Rocky Mountains and Andes Mountain ranges, with freshwater eventually returning into the Pacific via runoff (blue diagonal arrows). The purple thick line represents the approximate land mass coordinates of Central America used for computing zonal freshwater transport in the atmosphere.

comparisons of heat transport in CHIME and HadCM3, and 3). comparisons of zonal mean freshwater transport in CHIME and HadCM3 across the Panama Isthmus. A summary of the main points for each section is listed below:

**EPcm Model Heat Transport**

- analysed the model’s sensitivity to perturbations within the model’s two main parameters controlling the climate; the circulation strength parameter $K_A$, and the water vapour emissivity parameter $\gamma$, with default values of $100/15 Sv K^{-1}$, and 1.25 respectively.

- varied each parameter individually using multiples of their control values namely: 0.1, 0.2, 0.5, 1.0 (default value), 2.0, 5.0 and 10.0, both at control scenario $CO_2$ levels and for a doubling of $CO_2$.

- quantities measured were the surface temperature gradient $\Delta T$, the specific humidity gradient $\Delta q$, the atmospheric circulation strength $\Psi_a$, meridional latent heat transport
$L$, ocean heat transport $H_o$, atmospheric heat transport $H_a$, total heat transport $H_o + H_a$, and the ratio of heat transports $H_o/H_a$.

- observed increasing trends for both ocean and atmospheric heat transport with increasing values of the atmospheric circulation strength parameter $K_A$
- both increasing and decreasing trends to ocean and atmospheric heat transport with increasing $H_2O$ emissivity parameter $\gamma$. Maximum heat transport occurs for half the original value of $\gamma$.
- non-linearity of $\gamma$ to calculations of heat transport highlights very different responses of the climate when being in very cold, or very warm starting conditions.

**HadCM3 and CHIME Model Heat Transport**

- HadCM3 and CHIME are identical models except that the latter uses an isopycnal $z$ coordinate system within its ocean model, whilst the former uses constant depth; both models also use different mixing schemes to complement their choice of $z$ coordinate system.
- the comparison of both models allows a perfect test bed for ‘observing’ how sensitive climate models are to the type of vertical coordinate system it has for its ocean model.
- both models have almost identical distributions of total heat transport, constraining model differences for ocean and atmospheric components as of similar magnitude and different direction.
- CHIME has significantly less latent heat and dry static energy transport in the Northern Hemisphere Tropics compared to HadCM3 by approximately 1.2 and 1.4 PW respectively.
- reduction is the result of a weaker Hadley Circulation strength over the region by approximately 52%, even though low level tropospheric values of specific humidity increase by 9%.
- weakening of the Hadley Cell is hypothesised by the product of a flattening of the meridional tropical SST gradient in CHIME due to anomalously high SST values at Northern Hemisphere mid-latitudes.

The second section of this Chapter focused on the model comparisons of atmospheric freshwater transport between CHIME and HadCM3. A bullet summary is provided below:
HadCM3 and CHIME Atmospheric Freshwater Transport

- Both CHIME and HadCM3 have similar distributions of net vertical freshwater transport for each of the major basins and over land.

- Calculations with and without river runoff, the integrated areas of the Atlantic and Indo-Pacific basin are net evaporative regions, whilst the Southern Ocean and land precipitative.

- Calculations involving only net evaporation minus precipitation, model differences suggest that there is an anomalous flux of freshwater transport from the Atlantic into the Indo-Pacific basin in CHIME relative to HadCM3 of 0.05 Sv.

- Hypothesise that this atmospheric pathway is through the Panama Isthmus of Central America, where moisture is carried westwards by the Trade Winds.

- Under explicit calculations of freshwater transport over this region for the atmosphere, no we cannot prove this hypothesis under these conditions due to the associated errors being larger than the 0.05 Sv resolution needed to verify the claim. A larger perturbation is needed for any statistically significant conclusion.
Table 3.1: Graphical representation of all scenarios as a function of varying parameter value for both control and double CO$_2$ concentrations. The parameters include circulation strength $K_A$, emissivity of CO$_2$ $\alpha$, and the emissivity of water vapour $\gamma$. There are 42 different scenarios for analysis. Multiples represent by what value the original parameter value is multiplied by e.g. 0.1 for $\gamma$ represents a tenth of the original $\gamma$ value used. An additional 4 multiples (0.6, 0.7, 0.8, and 0.9) were done for $\gamma$, though these are not shown.

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<th>2.0</th>
<th>5.0</th>
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<th>0.1</th>
<th>0.2</th>
<th>0.5</th>
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<tr>
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</tr>
<tr>
<td>$\gamma$</td>
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<td>*</td>
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</table>
### Without Runoff

<table>
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<th>$F_{ip}$</th>
<th>$F_s$</th>
<th>$F_l$</th>
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</thead>
<tbody>
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<td>-0.76</td>
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<tr>
<td>HadCM3</td>
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<td>CHI-Had</td>
<td>0.04</td>
<td>-0.05</td>
<td>-0.01</td>
<td>0.03</td>
</tr>
</tbody>
</table>

Table 3.2: Net surface freshwater flux for the Atlantic (at), Indo-Pacific (ip), Southern Ocean (s), and land (l) contributions. Northward of $70^\circ N$ $F_S$ constitutes both land and sea areas. Lateral fluxes from continental runoff have been excluded. Data taken from years 80 - 119 of a control climate scenario. Both the Atlantic and Indo-Pacific regions are defined between latitudes $30^\circ S$ to $70^\circ N$, whilst the Southern Ocean is defined by the area southward of $30^\circ S$. Units are in Sv.

### With Runoff

<table>
<thead>
<tr>
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<th>$F_{ip}$</th>
<th>$F_s$</th>
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<td>CHI-Had</td>
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<td>-0.01</td>
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</table>

Table 3.3: Net surface freshwater flux for the Atlantic (at), Indo-Pacific (ip), Southern Ocean (s), and land (l) contributions. Northward of $70^\circ N$ $F_S$ constitutes both land and sea areas. Lateral fluxes from continental runoff have been included. Runoff values from the CHIME model have been modified to equal the net E-P rate over land. Data taken from years 80 - 119 of a control climate scenario. Both the Atlantic and Indo-Pacific regions are defined between latitudes $30^\circ S$ to $70^\circ N$, whilst the Southern Ocean is defined by the area southward of $30^\circ S$. Units are in Sv.
CHAPTER 4

Mass Transport in Temperature and Salinity space

4.1 Introduction

The classical ways of inferring motion in the ocean interior involves taking diagnostics of temperature and salinity, to be used in conjunction with the thermal wind and geostrophic equations for a reconstruction of interior flows. In particular zonal cross-sections of the ocean for inferred velocity are very useful for identifying major oceanic features such as the fast flowing western boundary currents of the Atlantic (Gulf Stream) and Pacific (Kuroshio), and also the deep return currents such as the North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW).

Intuitively, we have been brought up to understand structures in the standard 3 dimensional construct of xyz Cartesian coordinates, with its system for locating detail at specific spatial locations. Though this is indeed useful for the ocean, a zonal cross section of meridional velocity $V(x,z)$ lacks the clarity needed for differentiating water masses beyond the graphical separation of waters of the Atlantic and Pacific by continental structures. Waters at the same depth for both Atlantic and Pacific do not share the same properties of i.e. Temperature $T$ and salinity $S$, nor will waters of similar velocity $V$. Western boundary currents are highly concentrated jets of water that hug the coastlines with a diameter of around 100km across.
Although visible on an $V(x,z)$ plot the small spatial areas that these currents take up makes the study and identification of detail very difficult. Given these deficiencies a different approach has been borrowed that aims to address the problems of identifying and analysis water masses and their properties.

In this chapter we outline the basics for understanding water masses in temperature and salinity space. Firstly we describe present day theory on Water Mass Analysis (Walin, 1982; Stommel and Csanady, 1980). The second part analyses the water characteristics of the Atlantic and Indo-Pacific Oceans through meridional and zonal profiles of temperature, salinity and potential density. The last parts deal with the practical applications of this new procedure by the identification of signatures associated with the major Atlantic Water Masses. Analysis is conducted predominantly using the CHIME model with an associated comparison to the HadCM3.

### 4.2 Water Mass Analysis

The understanding of water masses in the ocean is best explained as analogous to the air masses of the atmosphere. This idea was developed during the 1920s in Norway’s Bergen School of Meteorology (Bergeron, 1959) whereby empirical observations evolved into scientific concepts.

Meteorologists normally describe 4 distinct atmospheric regions: Arctic/Antarctic, Polar, Tropical and Equatorial, with all being subdivided into regions of maritime (Open Ocean) and continental (over land) zones. Air masses formed within these regions are advected by local activity and form fronts when they meet, associated with strong winds and vertical motion.

Similar to the ocean, water masses formed at specific locations at the surface are advected below and into the ocean interior, forming oceanic fronts with strong gradients in temperature, salinity and density. Along with the atmosphere, the formation of fronts gives rise to mixing and diffusion of the two masses so as to homogenise the two media. Compared to the atmosphere where frontal systems can disappear after a few days, the ocean can take years to millennia for complete mixing to occur due to the slow nature of the ocean’s flow.

The analogy breaks down when we look at the defining properties of air (water) masses for the two media. For the atmosphere, an air mass is usually identified by its temperature and specific humidity. For most of the globe (except the equatorial upwelling region) an air parcel’s density is predominantly dependent on its temperature.
4.2.1 Water Mass Formation

Tomczak (1999) describes three main types of ways in which water masses are formed: convection, subduction, and sub-surface mixing.

Convection is a process whereby large fluxes of evaporative cooling increase the density of surface water, disrupting the stratification of the water column. The heavier water sinks to a depth until stability is reached and then advected throughout the ocean interior by local currents. Regions active to this type of convection include the Labrador Sea in the North Atlantic (Lilly et al., 1999; Lavender et al., 2002). To understand the processes of evaporative cooling we employ a simple theory described by Walin (1982) and Tziperman (1986).

In 1982 Walin proposed that the study of water parcels in temperature space could infer the same transport in physical space. Differential heating across the surface of the ocean led to density changes that in turn moved water between high and low latitudes analogous to an oceanic “Hadley Cell”. This type of circulation would be divided into a). poleward motion in the surface layer accompanied by cooling, b). sinking in the poleward region and equatorward motion accompanied by some mixing of water masses with slightly different properties, and c). motion towards the surface through the thermal stratification of the ocean, accompanied by heating through diffusion from above.

A consequence of this motion was that an individual water parcel would cycle through a range of temperatures depending on its position in the circulation, mapping out a motion in temperature space. Knowledge of this could therefore be inferred on the position of the water parcel within the circulation cell.

Extending this idea, Tziperman (1986) looked at two processes that acted to change the density of seawater: air-sea flux interactions, and mixing within the interior. A simple schematic for Tziperman’s ideas is reproduced in Fig. 4.1. We consider an ocean with two isopycnal surfaces $\rho_1$ and $\rho_2$ that outcrop towards the surface. Considering the effects of air-sea fluxes only, heat loss near the surface will result in densities of water $\rho < \rho_1$ being brought into the density range $\rho_1 < \rho < \rho_2$. Similarly, water parcels belonging to the latter density range is also exposed to the atmosphere, resulting in a cooling, but at a smaller rate to densities less than $\rho_1$. As a result, water at a density range between $\rho_1 < \rho < \rho_2$ will be cooled and be brought to a new density $\rho > \rho_2$. In this configuration more water is entering rather than leaving the density range $\rho_1 < \rho < \rho_2$ resulting in a source for this density type.

Concentrating now on small scale mixing, these processes aim to reduce difference in density between two water masses, leading to cross-isopycnal velocities. We consider a case where the
interior stratification leads to a net upwelling of water through the isopycnal $\rho_2$. Similarly if there is upwelling of water between $\rho_1 < \rho < \rho_2$ through the layer $\rho_1$ but at a stronger rate, this situation leads to more water leaving than entering then density range $\rho_1 < \rho < \rho_2$. Mixing processes thus act as a *sink* to this water type.

In steady state the summation of sources and sinks of a particular water type must equal to zero i.e. no net production. This constraint links the air-sea fluxes to the interior stratification of the ocean (Tziperman, 1986).

Subduction is predominantly a mechanical process whereby anti-cyclonic wind activity pushes surface waters down through isopycnal layers in what is known as *Ekman Pumping* (Stommel, 1979, see Fig. 4.2). This process is predominantly found in tropical ocean gyre regions with basin scale rotation being driven by westward (eastward) flowing winds near the equator (mid-latitudes). Similar small scale processes happen in ocean eddy activity which can vary in size between a few kilometres across to over 100 km (Stammer, 1998). The downward vertical velocities associated with this process are extremely slow, $\approx 30 m y^{-1}$ (Stommel, 1979). During the autumn and winter months the rate at which subduction occurs is overtaken by the growth of the mixed layer. Subducted waters are eventually convectively mixed with other waters, homogenising its properties where it is trapped for the rest of the winter months. On the onset of spring and summer a thin warm surface layer eventually separates the waters from the surface, and with Ekman Pumping, is forced downwards across isopycnal layers where it is advected away by local currents. Subduction has been found to be responsible for the formation of Central Waters (Marshall et al., 1993).

Sub-surface mixing is a rare event and the only one that does not depend on atmospheric
influences. This process occurs when two or more distinct water masses mix together so well that the resultant newly formed water mass has unique properties to that of the contributing waters. The best known example of water mass that is formed from this process is the Circumpolar Water found in the Southern Ocean (Thompson and Edwards, 1981), a region that is famous for the mixing of all the major ocean basin’s waters.

4.2.2 Formal Definitions of Water Mass Analysis

We consider a parameter space of \( n \) dimensions, each characterising a particular property of water. The major properties of water are temperature and salinity, so for this example the parameter space will have \( n = 2 \) i.e. a T/S curve. The parameter space does not have to be limited to two properties, and indeed other researchers do study other attributes such as oxygen concentration, nutrients etc.

We define a water type as a singular point in the \( n \)-dimensional space. Sticking with our \( n = 2 \) example of temperature a salinity, a water type would consist of a water mass with a unique temperature and salinity value. Usually in the ocean all water masses fill up distinct positions in the T-S space (see Fig. A).

A source water type describes water particles that develop properties very close to its formation region. Over time the region’s properties can change i.e. seasonally, leading to a spread of points in the \( n \)-dimensional space which have very similar properties to each other. Within this clustering we can define the mean water properties as being in the centre of this cluster in \( n \)-dimensional space, with the standard deviation being the diameter of the cluster in all \( n \) dimensional axes. Though this spread in the data constitutes a type of environmental variability, in practice there exists variability when measuring water properties from instruments such as CTDs, termed instrumental variability. These are on the order of 0.001\(^\circ\)C for temperature and 0.003 for salinity (Tomczak, 1999), with the majority of spread
due to environmental variability.

Water masses can be described as a source water type with corresponding standard deviations from its central position. Some water masses formed by convection have extremely well defined properties, requiring only a single water type and standard deviation to describe their space. These include Antarctic Bottom Water and North Atlantic Deep Water (Tomczak, 1999). Others are not so well defined, taking up a larger T-S spread. These include the Central Waters of the three main ocean basins and wind-mixed sub surface waters. In these examples it is easier to define these broad water masses by a collection of two or more central water types with corresponding standard deviations. A two point water type would be joined together forming a T-S mixing curve, a three point type would form a mixing triangle, and so on.

4.2.3 Water Mass Evolution after Formation

Newly formed water masses do not perfectly conserve their properties throughout their journey within the ocean; rather they evolve through certain stages. The first stage in which a freshly formed water mass goes through is what is known as water mass consolidation. In essence this process aims at reducing the spread, or standard deviation of water masses in its n-dimensional space. Instances where this occurs are in the formation region itself, with the formation of seasonally formed water masses mixing with the previous year’s water masses. This consequently reduces the environmental variance of the water mass by combining the ‘same’ water mass from different formation times. For a T-S curve, consolidation reduces the spread by converging on isopycnal surfaces, giving a smooth range of T-S properties for the same density class.

The second process undertaken is known as water mass ageing. As the description suggests, this is an absolute temporal change in the non-conservative properties of sea-water, relating predominantly to biological and chemical characteristics. This process happens to all water masses. In mathematical space, this would relate to a gradual drift in water mass properties over time.

The third process to occur is mixing between water masses. While this doesn’t change the physical properties of water masses, mixing results in a new sub set of water properties not found in water masses themselves. These sub sets would correspond to the T-S curve between two water types (2-D space), or the region encompassed by a mixing triangle (3-D space).

The above three processes described can be aptly grouped as water mass evolution. Given
enough time, eventually all water masses disappear from physical space through two processes: the first is known as absorption and describes the eventual absorption of smaller water masses into larger ones. An example is the absorption of Mediterranean Water into North Atlantic Deep Water. For other water masses, mixing with other water masses becomes so thorough that the original water properties disappears, only to be replaced by a completely new water mass. Both the latter processes describe the decay of water masses as an evolutionary process. Since the total volume of water is an almost conserved quantity (neglecting the effects of rapid ice melting, ice formation, evaporation and precipitation), all water masses are eventually recycled, with old waters forming the basis of new ones in the interior.

4.3 Water Mass Identification

The Continents and their geographical locations play a pivotal role in assigning certain ocean characteristics, with surface formation regions setting the water mass characteristics of the ocean interior. Due to their location, formation regions almost never have identical characteristics, making it possible to identify certain water masses belonging to certain regions. In the next few paragraphs we examine the three major ocean basins and describe each ones unique characteristics as a basis of water mass identification. These regions are the Atlantic, Pacific and Indian Oceans.

4.3.1 Atlantic

Figure 4.4 plots the latitudinal cross section of potential temperature, salinity and potential density anomaly $\sigma_t$ averaged between 300E (60W) to 360E (0W). This latter quantity is defined as the density difference of sea water from pure water, calculated as

$$\sigma_t = \rho - \rho_o$$

(4.1)

where $\rho_o = 1000 \text{kg} \text{m}^{-3}$. Data was taken from the CHIME model control run (ID 'cD_ocean') for the years 160-169. The warmest waters are found nearest to the surface and equatorial regions, decreasing with depth and proximity to the poles. The shape and distribution of the isotherms and isopycnals within the top 500m characterise the permanent thermocline and pycnocline respectively, with a large vertical change in temperature and density. Isopleths are not fully horizontal, bending towards the interior near the Tropics and approaching the surface within the equatorial region. These features are associated with Ekman vertical transport from
basin-scale gyres that pump water downwards in the former region and upwell them in the latter. High latitude waters in the Northern Hemisphere show weaker vertical stratification than their Southern Hemisphere counterparts. Indeed, temperature water properties extend to almost 2000m in the North, reflecting the processes of deep convection areas found at these high latitudes.

For salinity (Fig. 4.4B) the saltiest waters are found near the surface, particularly at subtropical latitudes where the downward Ekman pumping of water concentrates salinity values to sub-surface areas, forming salinity ‘lenses’. In terms of vertical size, these lenses are much larger in the Northern Hemisphere rather than the south, as a result of more intense gyre activity associated with superior parallel continental boundaries, and the near proximity to deep water formation regions that aids in downward vertical motion of water properties - these high sub-surface salinity waters are termed North Atlantic Common Water (NACW). At the southern latitudes between 50°S and 30°S, lies the region of the Southern Ocean, famously characterised as the only ocean to circumvent the entire globe. This area is also the formation region of the Atlantic’s freshest waters, commonly known as Antarctic Intermediate Water (AAIW). By tracing the shape of salinity isopleths that outcrop near the surface, AAIW sinks to a depth of between 300-1500m, where it is advected northwards up to ≈ 15°N, residing below saltier tropical NACW. For latitudes southward of 60°S lies the formation region of the densest major water mass in the Atlantic known as Antarctic Bottom Water (AABW). Being slightly saltier than AAIW but not as much as NACW, AABW is formed off the coast of Antarctica where brine rejection from ice formation, coupled with near freezing temperatures leads to very dense waters (Seidov and Haupt, 2001). These waters eventually slip off the Antarctic Shelf and flow northwards at depths below 4000m, making these particularly water masses the deepest flowing waters. The largest water mass (by volume) to exist in the Atlantic flows between AAIW and NADP at a depth of 1500-4000m, and is commonly known as North Atlantic Deep Water (NADW). This very dense water mass is formed from deep convection within the high latitudes of the Atlantic where it sinks and travels southwards. Work by Saenko et al. (2003) has shown that the interchanging of vertical positions for both NADW and AAIW are associated with ‘on-off’ states of the Thermocline. A summary of these mentioned water masses is given in Table 4.1 along with their property descriptions. As a comparison, Table 4.2 shows the same water masses but based on observational data taken from Marshall and Plumb (2007). All four Atlantic water masses represented in CHIME are approximately in range of the observed data. NADW encompasses a higher salinity upper limit (+0.4psu) compared
to observational data. The higher salinity lower limit on NACW is reflective of the model’s increased surface salinity bias, particularly for the Atlantic basin (Megann et al., 2010, Fig. 7c).

From our descriptions above, salinity offers an excellent diagnostic property in which to identify the major water masses in physical space. Since the contour shapes of these water masses are different in both temperature and salinity cross-sections, this ensures that each water mass has a unique signature on a T-S maps that are non-linear in space.

Density can be a useful proxy for depth in that lighter waters tend to remain on top of heavier waters.

Fig. 4.5A plots the meridional volume transport $V$ in Sv of the Atlantic as a function of $\sigma_t$ and latitude $y$ computed from

$$V(y, \sigma_t) = \frac{1}{\rho_0} \int_{x_1(y)}^{x_2(y)} \int_{\sigma_t - \frac{\delta \sigma_t}{2}}^{\sigma_t + \frac{\delta \sigma_t}{2}} vhdx'd\sigma'_t$$

where $\delta \sigma_t = 0.2$ is the potential density resolution, $v$ is meridional velocity, $x_1$ and $x_2$ are the longitudinal boundaries of the Atlantic basin respectively, $h$ is isopycnal layer thickness, and $\rho_0$ is a reference density. As depicted, light waters are transported northwards with the return flow being carried by heavier waters. Northward moving waters take up a very large range of densities (15-36 $\sigma_t$) characteristic of numerous surface and subsurface water masses between 0-1000m (Fig. 4.5B) that have been mixed together within the wind driven layers. The southward moving flow in comparison has a very restricted range of densities, featuring well defined interior water masses that contribute a stronger flow (in density coordinates) to its many northward moving counterparts.

From eq. 4.2 we construct the density integrated volume streamfunction $\Psi$ using

$$\Psi(y, \sigma_t) = \frac{1}{\rho_0} \int_{x_1(y)}^{x_2(y)} \int_0^{\sigma_t} vhdx'd\sigma'_t$$

with the Atlantic distribution of $\Psi(\theta, \sigma_t)$ shown in Fig. 4.5C, the convention of which positive (negative) values represent clockwise (anti-clockwise) circulations. The Atlantic possesses a dominant overturning circulation cell of roughly 15.7 $Sv$ that extends further south than the 34°S boundary imposed on our plots. Extensive research has been carried out on the variability of the Atlantic Meridional Overturning Circulation (AMOC - Bryden, 2005; Atkinson et al., 2010; Zhang, 2008) and its association to climate change (see section 1.5.5 for more details).
A lighter secondary cell is observed in the SH tropical region but at a smaller magnitude of 5.2 Sv.

We compute the average depth $D$ for which specific water densities reside in the Atlantic using the following equation,

$$D(y, \sigma_t) = \frac{1}{\rho_0} \int_{x_1(y)}^{x_2(y)} \int_{\sigma_t - \frac{\delta \sigma_t}{2}}^{\sigma_t + \frac{\delta \sigma_t}{2}} vhdx'd\sigma'_t$$

$$= \frac{1}{\rho_0} \int_{x_1(y)}^{x_2(y)} \int_{\sigma_t - \frac{\delta \sigma_t}{2}}^{\sigma_t + \frac{\delta \sigma_t}{2}} vdx'd\sigma'_t$$

(4.4)

The computation of $D$ enables a mapping of $\Psi$ from density coordinates to vertical depth coordinates, represented in Fig. 4.5D. The large overturning circulation extends from the surface to nearly 3000m below, being primarily composed of NACW, NADW and AAIW. The smaller secondary circulation, although still quite prominent in density space, only resides in the upper 150m of the ocean, clearly dwarfed by the AMOC in physical space.
### CHIME Model

<table>
<thead>
<tr>
<th>Water Mass</th>
<th>Temperature °C</th>
<th>Salinity psu</th>
<th>$\sigma_t$ kgm$^{-3}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>NACW</td>
<td>10+</td>
<td>35.7+</td>
<td>31.8 - 35.8</td>
</tr>
<tr>
<td>AAIW</td>
<td>5 - 6</td>
<td>34.3 - 34.7</td>
<td>35.8 - 36.6</td>
</tr>
<tr>
<td>NADW</td>
<td>2 - 4</td>
<td>34.9 - 35.4</td>
<td>36.9 - 37.2</td>
</tr>
<tr>
<td>AABW</td>
<td>-0.1 - 1.3</td>
<td>34.5 - 34.7</td>
<td>37.0 - 37.2</td>
</tr>
</tbody>
</table>

Table 4.1: Property definitions of the major water masses found in the Atlantic basin, empirically derived from CHIME’s control run, years 160-169.

### Observational Data

<table>
<thead>
<tr>
<th>Water Mass</th>
<th>Temperature °C</th>
<th>Salinity psu</th>
<th>$\sigma_t$ kgm$^{-3}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>NACW</td>
<td>10+</td>
<td>35+</td>
<td>-</td>
</tr>
<tr>
<td>AAIW</td>
<td>4 - 8</td>
<td>34.3 - 34.9</td>
<td>-</td>
</tr>
<tr>
<td>NADW</td>
<td>2 - 4</td>
<td>34.9 - 35.0</td>
<td>-</td>
</tr>
<tr>
<td>AABW</td>
<td>-0.1 - 2.0</td>
<td>34.7 - 34.8</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 4.2: Property definitions of the major water masses found in the Atlantic basin, empirically derived from observational data. Values based on data taken from Marshall and Plumb (2008).
4.3.2 Indo-Pacific

The Pacific is the largest of the ocean basins, stretching from the western coastlines of the Americas on the East, to South-East Asia on its left. Its maximum northward extent stretches to approximately $60^\circ N$ where the narrow channel of the Bering Strait separates it from the Arctic. At its southern most extent, there is no fixed latitude boundary, but meanders south of Australia and the tip of South America, northward of the Southern Ocean. Its sheer size is a hub of activity for both ocean and atmosphere processes. Slow oscillations of warm and cold waters near the equator drive common atmosphere events such as El Nino and La Nina. The summer monsoon seasons provide a major source of freshwater to large parts of East Asia, along with large scale weather patterns in the form of typhoons.

Properties of temperature, salinity, and density for the Pacific are plotted in Fig. 4.6 taken from CHIME. We note that the color scaling for Fig. 4.4, Fig. 4.6 and Fig. 4.7, are not identical since each region has a different range in values.

Many of the Pacific’s properties are symmetric about the equator. Unlike the Atlantic, there is no weakened stratification of the water column at high northern latitudes, with the permanent thermocline situated between 300-500m. The continental geography bounding the Pacific limits the northern extent of the Pacific at the Bering Strait of approximately $60^\circ N$. Along with the increased precipitation found within the North Pacific, these inhibit any deep water formation processes found at higher latitudes in the Atlantic (Warren, 1983; Emile-Gaey, 2003). Within the Southern Hemisphere at latitudes between $50^\circ S$ and $30^\circ S$, the temperature of the water column remains very uniform up to 1000m, consistent with the formation region of AAIW.

A marked contrast for the Pacific region is that it is considerably fresher throughout the entire column compared with the Atlantic (another feature inhibiting deep water formation processes).

The density cross section remains remarkably similar to that of the Atlantic given the different features in temperature and salinity.

The Indian Ocean is the smallest of all the major ocean basins, bounded by the east coast of continental Africa on its left, and the Islands of Indonesia, Malaysia and the Philippines to its right. It has the smallest latitudinal extent, reaching approximately $5^\circ N$ at its north, to the edge of the Southern Ocean region below, at around $40^\circ S$. Cross sectional areas of temperature, salinity and density for this region is shown in Fig. 4.7 with the same data set used with the Atlantic and Pacific regions. The longitudinal boundaries are situated at $40^\circ E$. 
and 110°E. The features for all three properties are again very similar to that of the Atlantic and Pacific, though with salinity being closer matched to the latter region. AAIW formation is visible in this region though the northward extent of its reach is limited by geography to just south of the equator. Although the saltiest waters lie near the surface, CHIME exhibits an extra surface layer of fresh water on top of this salty layer, representing regional continental runoff from the Indian and Pakistan subcontinents.
The Pacific and Indian Oceans are both unique regions each with their own characteristic features. However, unlike the Atlantic, the borders between the two regions are extremely hard to define given the broken passages of South East Asian islands. Due to these difficulties it is normal practice to group together the Pacific and the Indian Oceans (Indo-Pacific, abbreviated to IP) as a single entity, particularly for the computation of overturning streamfunctions that need ‘concrete’ longitudinal boundaries. Fig. 4.8 plots $\Psi$ integrated across both ocean basins. Meridional volume transports share similar ranges in density with the Atlantic with light poleward waters being adected near the surface and denser return flow water underneath (Fig. 4.8C). In density coordinates the IP region exhibits two symmetric overturning cells on either side of the equator, occupying a large density range of between 30-35 $\sigma_t$, larger than that for the Atlantic. In physical space however these cells only occupy the top 300m of the ocean (Fig. 4.8D) and are associated with wind driven gyre circulations.

We briefly conclude the main characteristics of the major ocean basins being,

- Atlantic is considerably saltier than all other basins.
- Atlantic circulation is dominated by a single overturning cell extending into the Southern Hemisphere
- Indo-Pacific region has fresher waters
- Indo-Pacific circulation is dominated by shallower overturning cells on either side of the equator
- All basins share Antarctic Intermediate Water and Antarctic Bottom Water characteristics that are advected from the Southern Ocean

4.4 $M(T,S)$ three dimensional analysis

The classical two dimensional analysis of water masses using a T-S diagram, although useful, lacks any information on the relative importance of individual water type ‘points’ in terms of the magnitude of transport. This section merges together the T-S map analysis with meridional mass transport $M$ for individual water types into what will be termed $M(T,S)$. Model data is used to provide the necessary spatial and temporal resolution needed for this type of analysis that is not found in observational data sets.

The data used has been taken from the CHIME model that has an isopycnal based vertical coordinate system. This choice makes this model ideal for water mass analysis as it has been
shown to preserve water mass properties (both temporal and spatial) much better than the majority of models that use constant z-depth for their vertical coordinate choice (Megann et al., 2010).

In the next few sections we define the major characteristics of the M(T,S) diagrams, concentrating on the main features associated with Atlantic and Indo-Pacific basins.

Since the vertical height associated with individual water masses are not shown on M(T,S) diagrams, we show how we can use (to a first order approximation) potential temperature as a form of z coordinate to analyse the characteristics of near surface waters to those within the interior of the ocean. Although there are seasonal effects on M(T,S) maps, particularly for near surface waters, we concentrate predominantly on the time mean characteristics. As the number of zonal cross sections available are too numerous to discuss them all in detail, a select few are chosen that encompasses the majority of the ocean’s major water masses. These sections are: 40°N, 24°N, 0°N, 20°S, and 34°S.

### 4.4.1 M(T,S) basics

An M(T,S) diagram is essentially an extension of the classical T-S diagram but with the addition of an extra component in the z-component for each individual mass transport. A setup of a T-S diagram is shown in Fig. 4.9 with salinity (temperature) on the x (y) axis respectively. The coloured curved lines are contours of constant potential density anomalies $\sigma_t$, or isopycnal surfaces, calculated from Eq. 4.1

Potential Density $\rho$ is a highly non-linear variable that changes with temperature, salinity, and pressure such that

$$\rho(\theta, S, P)$$

Density is proportional to salinity and pressure, but inversely proportional to potential temperature; salinity adds mass to seawater by increasing the dilution concentration of salt within a fixed volume of water, whilst increases (decreases) in potential temperature lead to an expansion (contraction) of seawater and a lower (higher) density. Pressure tends to squash water together making it denser and thus heavier.

Although pressure does increase with depth in the ocean, seawater is extremely hard to compress, unlike that for atmospheric air parcels. It is sometimes general practice to omit the dependency on pressure by assigning a constant value when calculating density. For CHIME, ocean pressure is set at a constant value of 2000 dbar (1 dbar = 10 bar = 1,000,000 Pa), resulting in a dual density dependence on temperature and salinity only. Even with the use
of this simplification, density is still a highly non-linear function illustrated by the curved isopycnal contours in Fig. 4.9. At ‘warm’ temperatures ($T > 15^\circ C$), changes to density couple much stronger to changes in temperature rather than salinity. This can be explained by the distance between adjacent contours in the x and y direction, where it is shortest in the T axis rather than the S axis. Below this temperature the density contours start to become non-linear with an increase in contour curvature. In this temperature range the shortest distance to an adjacent density surface is now in the S axis as opposed to the T axis. Thus at low temperatures, changes to density couples stronger to changes in salinity.

The addition of meridional mass transport on the T-S map is represented by individual water types, or dots on the map, with colour indicating magnitude and direction. The direction of magnitude is on the z direction of the T-S plot, giving a 3-D representation of water masses. The resolution of our $M(T,S)$ plots is $\delta T = 0.2^\circ C$ and $\delta S = 0.1 psu$. Roughly scaling a sensitivity to temperature and salinity of 1/100 and 1/10000 respectively, would imply a density sensitivity of 1/100,000, or approximately resolved differences of 0.1 $\sigma$. $M(T,S)$ plots thus have the capabilities of resolving very fine detail water masses.

We calculate $M(T,S)$ using the equation

$$M(T, S) = \int \int \rho v \frac{1}{\delta T \delta S} dx dz$$

(4.6)

where $v$ is meridional velocity, $dx$ and $dz$ are zonal and vertical increments respectively; all other symbols have their usual meaning.

The solid horizontal lines and dashed vertical lines represent the area weighted mean temperature and salinity values for poleward moving waters, the colour convention being blue (red) for northward (southward) moving waters. These are calculated from

$$Q_{(N,S)} = \frac{\int_{(0,-\infty)}^{(\infty,0)} MQdM}{\int_{(0,-\infty)}^{(\infty,0)} MdM}$$

(4.7)

where $Q$ represents our are weighted quantity i.e. temperature or salinity, and $M$ is the meridional mass transport for a water type computed as the product of meridional velocity and zonal area. The subscripts $N$ and $S$ represent northward and southward moving water property values respectively, with the convention that all positive values travel northward. Eq 4.7 essentially finds the ‘centre of mass’ values in the temperature and salinity axis. We will see later that although there can be large distribution of water types scattered in T-S space, the differences between area weighted mean values for northward and southward moving waters can be very small compared to the extreme ranges in temperature and salinity.
Using $24^\circ N$ as an example, we plot the corresponding $M(T,S)$ plot on Fig. 4.10. Units of mass transport are in $Sv \times \rho$ ($1 \text{ Sv} = 1 \times 10^6 \text{ m}^3 \text{ s}^{-1}$). Each of the four sub-figures are identical except that the colorbar magnitude for each successive plot reduces, increasing the sensitivity of the $M(T,S)$ plots. In Fig. 4.10A details of mass transport are limited to densities greater than $\sigma_t > 33.1 \text{ km}^{-3}$. At resolutions less than a fifth of the maximum range of $M$ (Fig. 4.10C, and D), details of water masses for $\sigma_t < 33.1 \text{ km}^{-3}$ become visible, differing in both magnitude and spatial distribution to its denser waters. We remark that these smaller magnitude lighter waters represent near surface wind driven circulations, whilst the denser waters involve interior flows - more details are described below. We note how water masses tend to cluster around lines of constant $\sigma_t$, mirroring the isopycnal coordinate system of the CHIME ocean model. As a convention, all further $M(T,S)$ plots use a magnitude sensitivity of $\pm 0.05 \text{ Sv}$ to include details of these near surface circulation behaviours.

To aid in our explanation of our example $M(T,S)$ the corresponding zonal cross sections of $T$, $S$, $M$, and $\rho$ for $24^\circ N$ are included in Fig. 4.11.

To understand the basic details of our example $M(T,S)$ map at $24^\circ N$ we plot the corresponding zonal cross-sections of potential temperature, salinity, meridional mass transport, and density anomalies in Fig. 4.11.

Ocean temperatures can reach up to $30^\circ C$ in some geographical locations near the surface, with the vast majority of the ocean remaining below $10^\circ C$. The ocean warms near the surface via absorption of solar radiation, which is then filtered down into the ocean interior by a combination of mechanically induced mixing (winds and tides) and diffusion. Temperature decreases with depth and as such can be used as an approximate proxy for vertical depth (Fig 4.11a).

Ocean salinity (Fig. 4.11b) is directly affected by the net input/output of freshwater with the main moderator being the atmosphere; precipitation leads to a dilution of seawater, reducing salinity whilst evaporation enhances it. Other sources of freshwater are more land-based and involve freshwater river discharges (land run-off), and the melting of glacial ice. Though found in Polar Regions, the formation of ice due to freezing leads to brine rejection of seawater, increasing salinity. This is an important mechanism near the Antarctic region where heavy continental waters spill over the shelf and sink into the ocean interior as Antarctic Bottom Water (Foster and Karmack, 1976; Saenko and Weaver, 2001). Near the surface, salinity is non-spatially distributed with the Atlantic being considerably saltier than its Indo-Pacific counterpart.
Meridional mass transport is an extremely useful property for understanding the internal circulation of the ocean, for its spatial distribution holds vital information to certain mechanisms. The graphical representation of mass transport in Fig. 4.11c has had its colorbar rescaled by a decrease of one order of magnitude to intensify the details of fine structures. Within the top 600m the ocean dynamics are predominantly wind-driven that include basin-scale lateral gyre circulation. This is characterised by a very thin, intense northward flow on the western side of the basin, with a slower but broader scale return flow on the eastern side (Stommel, 1948). These Western Boundary Currents are some of the fastest flowing waters found in the ocean and are well known in the North Atlantic (Gulf Stream) and Pacific (Kuroshio). In Fig. 4.11c the western intensification at 24° N travels off the coast of Eastern China in the Pacific and around the Gulf of Mexico in the Atlantic. For the latter region this water current remains in tight proximity to the coast. Below 2000m southward transport of water occurs completing an overturning circulation in the vertical that is synonymous with the large scale Hadley Cells in the atmosphere. Vertical variations in meridional mass transport are greater in the Atlantic than Indo-Pacific region highlighting the richness of activities and mechanisms in the former, even though it is of smaller size (see Fig. 4.5d and 4.8d).

As noted above the relationship between $\rho$, T and S is non-linear, with changes in density coupling stronger to T at high temperatures (approximately $T > 5^\circ C$) and to S for low temperatures (approximately $T < 5^\circ C$). The cross section of $\sigma$ thus resembles a similar distribution to potential temperature above 600m. We note that for the Atlantic basin the layers of density are not as stratified as for the Indo-Pacific shown in potential temperature, since the Atlantic has relatively higher salinity values above the thermocline, increasing the density of these waters. Below 600m density properties are highly uniform with small changes a consequence of variations in salinity.

### 4.4.2 M(T,S) - Atlantic and Indo-Pacific basin Characteristics

Figure 4.12 plots M(T,S) maps for the global cross section at 24° N. From the left, the three vertical columns represent the global mean oceans, the Indo-Pacific Basin, and the Atlantic Basin respectively. From the top, the three horizontal rows represent the net flow, the northward component, and the southward component respectively. The units are mass transport, given as $\rho \times Sv/\delta T \delta S$. / 

The global ocean at 24° N is characterised by two large scale T-S curves that have similar temperature, but different salinity ranges. These curves represent the Indo-Pacific (Fig. 4.12d)
and Atlantic basins (Fig. 4.12g), showing very distinct characteristics that take up individual and highly separable locations in T-S space. The Atlantic basin is predominantly characterised by higher salinity waters throughout the whole temperature range, whilst the Indo-Pacific is distinctly fresher. This feature is mostly exaggerated at the higher end of the temperature range closest to the surface where the modulation of water mass properties is greatly influenced by atmospheric processes. At the lower end of the temperature range, the M(T,S) diagrams samples interior water masses, with water properties from both basins starting to convergence. Within these regions, differences between water masses are not as large near the surface due to the processes of water mass evolution, consolidating water masses.

The distributions of northward moving waters (Fig. 4.12b, e and h) have a tendency to include the warmest water properties, increasing the mass weighted temperature of northward moving waters. Compared with the distributions of southward moving waters, (Fig. 4.12c, f, and i), these water masses tend to reside at lower temperatures, thus having a lower mass weighted temperature. These observations are consistent with an overall northward heat transport (at this latitude) with warm waters travelling northward near the surface coupled to a cold return flow below. The same analogy is true for mass weighted salinity values, with salty water travelling northward due to excess evaporation at the surface, and fresher return water below.

For mass weighted northward moving temperatures (Fig. 4.12b, e and h) the Atlantic has a significantly higher temperature compared to the Indo-Pacific or global mean. In contrast to southward moving temperatures (Fig. 4.12c, f and i) that remain similar for both regions, imply that the Atlantic has a much larger temperature gradient. From estimates of heat transport at 24°N (Trenberth and Caron, 2001) both regions transport roughly equal amounts of heat transport northward. Since heat transport is proportional to the temperature gradient and meridional mass transport, these results suggest that much of the Atlantic basin’s heat transport has a stronger dependency on its temperature gradient, compared to the Indo-Pacific which relies more on mass transport.

The components of overturning and gyre components for the ocean circulation can be inferred from information on M(T,S) diagrams. As an example we shall use plots from the Atlantic basin i.e. Fig. 4.12g, h and i. Gyre circulation is predominantly confined to near surface lateral movements, most commonly linked to the strong and narrow northward flow of the western boundary Gulf Stream being balanced by the weaker and broader flow on the eastern side. As a rough guide this would describe all water mass features above a temperature
of approximately 20°C. For the overturning component, the near surface northward flow of warm waters are balanced by the equatorward transport of colder waters at depth. Water masses below 20°C can thus be used to segment this mechanism. As already shown, the Atlantic exhibits a predominantly overturning circulation component, whilst the Pacific is more gyre orientated.

### 4.5 Applications

Having covered the theory on water masses and the basics of $M(T,S)$ diagrams we apply our diagnostics to a specific number of latitudes that cover the majority of the Atlantic region: 34°S, 20°S, 0°N, 24°N, and 40°N. These are the same latitudinal cross-sections shown in Fig. 4.4 as the black vertical dashed lines. The analysis aims to show specific signatures for the major water masses in the Atlantic, and how the distributions in $M(T,S)$ space evolve with latitude.

Figure 4.13 plots the $M(T,S)$ diagrams for the latitudes discussed previously with each row constituting a unique latitude. The global zonal integral of water masses are shown on the left column whilst only waters from the Atlantic region are shown on the right. The boxed regions represent the T-S space unique to certain types of water masses encompassing: NACW (green), AAIW (red), NADW (blue), and AABW (black). Each boxed T-S limits are defined empirically for individual latitudes. Data is taken from the transient CO$_2$ scenario for years 85-94, representing a CO$_2$ concentration of $1.4 \times$ pre industrial values, or 380 ppm (present day concentrations).

For the global analysis (Fig. 4.13A, C, E, G, I) there are unique characteristics and evolutions of the $M(T,S)$ diagram for each latitudes. The circulation of warm poleward and cold equatorward moving waters are still prevalent in both hemispheres. The colour change (blue above red) for Southern Hemisphere latitudes is still consistent with this circulation since positive $M$ values correspond to northward moving waters across both hemispheres. For the Atlantic basin there is no colour change across the equator, i.e. northward moving water is always warming than southward moving water across the entire basin, reflective of the cross equatorial AMOC.

For Southern Hemisphere latitudes, (34°S and 20°S) the $M(T,S)$ branch structures for the Atlantic and Indo-Pacific basins become very difficult to differentiate, particularly for interior waters of $\sigma > 35.1 kgm^{-3}$. At these latitudes, the close proximity to the Southern Ocean ($\theta < 34°S$) mixes waters from all basins, homogenising water mass properties. Only
the lighter signatures from near surface water masses have differentiated signatures. Further northwards into latitudes $0^\circ N$, $24^\circ N$, and $40^\circ N$, basin signatures start to diverge and occupy non-overlapping T-S spaces, leading to the classic characteristics of a fresher Indo-Pacific and saltier Atlantic. At these latitudes the decreased influence of the Southern Ocean gives the ability for both regions to develop their own individual water mass characteristics from different geographical formation regions and continental boundary effects. Indeed, the differences between northward and southward moving water mass temperatures (horizontal red and blue lines) are larger for northern latitudes, furthest away from the Southern Ocean. For the Southern Hemisphere, the almost similar temperatures again add evidence to a greater mixed region of water.

### 4.5.1 Atlantic Water Mass Signatures

The Atlantic is made up of four distinct water masses (see Table. 4.1) that occupy unique positions on the T-S map, highlighted by the coloured boxes in Fig. 4.13. Both AAIW and AABW do not penetrate further than $24^\circ N$ and thus do not have individual boxes for latitudes greater than this. The spatial patterns of all Atlantic interior water masses, that is to say, AAIW, NADP and AABW are found in relatively similar locations throughout all computed latitudes south of $24^\circ N$. AABW is the easiest water mass to identify; being the deepest flowing, it is therefore the densest and coldest water mass, having its M(T,S) signature the lowest in T space (black boxes). NADW is identified as an intense cluster of water types just north-west of the position for AABW, being slightly saltier and warmer. Just like AABW, its position in T-S space is very localised reflecting almost uniform water properties throughout all five specified latitude cross-sections. Being the freshest water masses out of the four, AAIW is characterised by a narrow T-S space that includes a large group of the least saline water types flowing northward. The large temperature range (greater than the stated $5 - 6^\circ C$ in Table 4.1) is due to the unforeseen isopycnic water types of similar salinities that extended further out of this temperature range. At some latitudes, namely $34^\circ S$ and $0^\circ N$, the upper temperature limit for AAIW is very difficult to quantify since clear isopycnal water mass signatures of AAIW mix in with near surface wind driven ‘noisy’ signatures (see Fig. 4.4). The upper temperature limit at these latitudes should therefore not be taken literally. Lastly, NACW occupies much of the sub-tropical near surface waters, and so is prevalent to atmospheric dynamics of evaporation and wind stress. The NACW thus has characteristics of being the warmest and saltiest water masses. Due to its near surface proximity, both temperature and salinity ranges for NACW
are also the largest for any of the four major Atlantic water masses. Water types are not grouped onto individual isopycnal surfaces such as AAIW, but encompass a continuous range, usually for $\sigma_t > 35.1$.

For the northern hemisphere latitudes (Fig. 4.13H, J) signatures of AAIW and AABW disappear, leaving only an Atlantic make-up of NADW and NACW. NACW become increasingly dominant on the M(T,S) now encompassing signatures of noisy wind induced water types, but also clear isopycnal flows. For the latter description, NACW at these latitudes are not segregated to the top 200m of the water column but extend to depths of up to 2000m below. It is at these depths where the surface wind-induced effects are not felt do water types flow on isopycnal surfaces (see Fig. 4.4B).

The temperature difference between northward and southward moving water masses for the Atlantic far surpasses the global average, in particular for latitude northward of the equator. This feature compliments the single AMOC configuration of the Atlantic basin where all upper surface waters travel northward compensated by deeper and colder return flows. As a side note, the large temperature difference at 24°N agrees well with a near maximum of poleward heat transport for the Atlantic basin.

Analysis on M(T,S) graphs allows for a much better tool to giving the ranges of T-S properties for individual water masses as locations are clearly distinct. The differences in values found in Table 4.1 are also due to the different methodologies in calculating each water mass property range. Whereas Table. 4.1 uses a continuous latitudinal profile but for a limited longitudinal average (60°W - 0°N), the M(T,S) method only analyses one specific latitude at a time and with full Atlantic basin coverage (in the longitudinal direction).

We expect that the salinity and temperature ranges for the different Atlantic water masses to be subtly different at each latitude due to the mechanisms of water mass evolution (consolidation and mixing). Whilst we do not analyse oxygen or nutrient concentration as a proxy to the age of water, the break-away of AAIW from NACW shows water mass consolidation in practice, particularly at 20°S and 0°N.

In retrospect to our previous identification of water masses from cross-sections of salinity, temperature offers a surprisingly good variable for identifying and separating individual water masses on a T-S map.

In later chapters we analyse how water mass signatures at these latitudes evolve in a ‘global warming scenario’.
4.5.2 Meridional Ocean Heat Transport

The M(T,S) plot provides all the natural tools, albeit a more time-consuming method, for calculating the meridional heat transport of the ocean. We calculate the latter from

\[ H_o = c_o M_o \Delta T \]  \hspace{1cm} (4.8)

where \( c_o = 4000\text{Jkg}^{-1}\text{K}^{-1} \) is the specific heat capacity of seawater, \( M_o \) is the total poleward or equatorward transport meridional mass transport, and \( \Delta T \) is the temperature difference between mass weighted poleward and equatorward moving water.

Meridional Mass transport \( M_o \) is calculated as a summation of all positive or negative mass transports, given by

\[ M_o^+ = \Sigma \rho_i v_i^+ dA_i \]  \hspace{1cm} (4.9a)

\[ M_o^- = \Sigma \rho_i v_i^- dA_i \]  \hspace{1cm} (4.9b)

\[ M_o^+ = |M_o^-| = M_o \]  \hspace{1cm} (4.9c)

where the superscripts of ‘+’ and ‘-’ refer to the summation of all positive or negative values of velocity \( v \) and density \( \rho \), and \( dA_i \) represents the corresponding zonal area for each individual velocity bin.

Mass weighted northward moving temperature values are calculated using eq. 4.7, which in summation form is,

\[ Q(N,S) = \frac{\Sigma_{i=(+\infty,0)} v_i T_i dA_i}{\Sigma_{i=(0,-\infty)} v_i dA_i} \]  \hspace{1cm} (4.10)

The temperature difference, or temperature gradient is thus defined as \( \Delta T = T_N - T_S \).

This convention is better than stating ‘poleward’ or ‘equatorward’ moving water masses since \( \Delta T \) must be negative in the southern hemisphere to accommodate for a southward moving heat transport; values that cannot be attained using the latter definitions.

A complete analysis of the latitudinal distributions for the various quantities in eq. 5.3 is shown in Chapter 3,5 and 6 where both data from CHIME and HadCM3 is used, again highlighting the versatility of M(T,S) plots.
4.5.3 Ocean Freshwater Transport

The use of $M(T,S)$ plots for the ocean can be used to calculate the meridional freshwater transport of the ocean. Calculations of the salinity difference based on Eq. 4.10 are used with the variable ‘$Q$’ being replaced with salinity $S$. To avoid the duplication of equations with our derivations a complete analysis of this section is covered in the next chapter.

4.6 Comparisons to HadCM3

CHIME’s vertical isopycnal coordinate system for its ocean component (coupled with its KPP mixing scheme (Large et al., 1994)) allows for a superior representation of water masses in physical space. Its coordinate system is naturally complimentary to $M(T,S)$ maps, with interior water masses lying perfectly on surfaces of constant $\delta_r$. As a comparison, the HadCM3 model employs a constant depth vertical coordinate system with corresponding Bulk formulae mixing scheme (Gordon et al., 2000). We expect interior water mass features to be less well-defined on isopycnal surfaces, having greatly diffuse characteristics on $M(T,S)$ maps in particular, AAIW (Megann et al., 2010).

Figure 4.14 plots identical maps of $M(T,S)$ calculated in CHIME (Fig. 4.13) but now for HadCM3. Data is taken from a control run scenario for the years 80-119 from spin-up. The method for generating $M(T,S)$ plots for both models are identical. We note a change in magnitude scaling, with HadCM3 using limits of $\pm 0.1 \, \rho Sv/\delta T \delta S$ as compared to the $\pm 0.05 \, \rho Sv/\delta T \delta S$ used in CHIME. The change was necessary since HadCM3 exhibited substantially stronger mass transports than CHIME, warranting the rescaling to increase detail.

Overall the maps of $M(T,S)$ for HadCM3 share the large scale features shown in CHIME: the diverging of water mass features for northern hemisphere latitudes for the Atlantic and Indo-Pacific basins, the positive Atlantic temperature gradient for all specified latitudes etc.

Just as hypothesised above, the majority of interior water masses for $\theta > 34.1$ are extremely diffuse on isopycnal surfaces, with the majority of water types having a more continuous range of densities rather than discrete values. The exception to this detail however is at the equator (Fig. 4.14E and F), where substantial water types do localise onto discrete densities, again characteristic of water mass consolidation for advected AAIW from the Southern Ocean.

For the very diffuse water masses related to near surface wind driven mechanisms, the spread in T-S space is much more localised for HadCM3, offering a smaller range in water mass properties compared with CHIME.
Within the Atlantic the T-S curve for interior water masses share similar characteristics to that of CHIME with both AABW and NADW holding almost identical positions. The identification of AAIW is much more difficult since its water mass characteristics are extremely diffuse, merging with NACW features. Again, this observation is consistent with a very diffuse AAIW (in physical space) after 80 years of integration, compared to a better maintained water mass structure in CHIME for the same period (Megann et al., 2010).

Comparing M(T,S) at 40°N for both models (Fig. 4.13I and Fig. 4.14I) there is evidence for substantial differences in water mass properties in the Indo-Pacific basin. Whilst CHIME may be described as having its water masses scattered around a large T-S area, HadCM3 in comparison has very linear water mass characteristics. In particular HadCM3 is observed to have an extremely cold and fresh southward flowing water mass that is not seen in CHIME for this region.

4.7 Summary

Just like the atmosphere, the ocean has a very complex internal structure comprising numerous large scale water mass flows situated at specific heights. The majority of water masses attain their properties near the surface before sinking and being advected internally. Through its journey within the ocean, processes happen whereby individual water masses consolidate their properties onto isopycnal surfaces, or mix together with adjacent water masses to form new ones. The formation of water masses is a continuous process with the extinction of one leading to the creation of another.

M(T,S) maps are a highly versatile tool that stretches far beyond mapping water masses onto T-S space. The identification of certain water masses and their characteristics can sometimes be substantially more accurate than through empirical identification in physical space. In terms of overall basin scale characteristics, they allow for the easy identification of the salty Atlantic and fresher Indo-Pacific basins. Sharply defined isopycnal water masses of the ocean interior are contrasted to the weaker, diffuse characteristics of near surface waters from mechanical mixing and air-sea interactions.

Our direct comparisons of M(T,S) maps from CHIME and HadCM3 has allowed (possibly the first of its kind) a way of understanding how switching from a constant z to a constant isopycnal vertical coordinate system in the ocean can affect water mass properties. Although both models share the similar large scale features associated with individual basins and their corresponding water masses, the patterns of water mass distribution are much diffuse around
isopycnal surfaces for the constant z depth model HadCM3, and at some latitudes include new water masses altogether.

Water mass analysis contains a holistic approach to the ocean, whether it be water properties, mass transport or water mass evolution. On a more relevant note, the present associations with climate change will extend this type of analysis to data taken for predicted oceanic climates under anthropogenic forcing.
Figure 4.3: a). T-S curves for the North Atlantic and North Pacific at 24°C, along with meridional cross sections spanning subtropical and subpolar regions, b). magnified version for the deep waters. All data taken from WOCE. Figure taken from Talley (2008, fig. 8).
Figure 4.4: Latitudinal cross-sections of A. Potential temperature, B. Salinity, and C. Density Anomalies $\sigma_t$ for the Atlantic Basin averaged between 60W and 0W. The Salinity cross-section includes the approximate positions of the major Atlantic Water masses. The dashed black lines represent specific latitudes of further analysis. Data taken from CHIME control run, years 160-169.

Figure 4.5: A). Atlantic Ocean volume transport on $\sigma_t$ levels, B). depth of $\sigma_t$ levels, C). meridional overturning streamfunction in $\sigma$ coordinates, D). meridional overturning streamfunction in depth coordinates. Data taken from CHIME control run, years 160-169.
Figure 4.6: Similar to Fig. 4.4 but for the Pacific Basin, bounded by the longitudinal coordinates 150E to 250E.

Figure 4.7: Similar to Fig. 4.4 but for the Indian Ocean, bounded by the longitudinal boundaries 40E to 110E.
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Figure 4.8: Indo-Pacific Ocean volume transport on $\sigma_t$ levels, B). depth of $\sigma_t$ levels, C). meridional overturning streamfunction in $\sigma$ coordinates, D). meridional overturning streamfunction in depth coordinates. Data taken from CHIME control run, years 160-169.

Figure 4.9: Map of T-S space. Curved coloured lines indicate surfaces of constant $\sigma_t$ computed at a constant pressure of 2000 dbar.
Figure 4.10: $M(T,S)$ at 24N for CHIME TRAN scenario, years 85-94. All four figures are identical except for rescaled colorbars of volume transport, set to A). $\pm 0.5$, B). $\pm 0.25$, C). $\pm 0.1$, and D). $\pm 0.05$.

Figure 4.11: Zonal Cross-sections at 24°N for A). potential temperature, B). Salinity, C). meridional mass transport, and D). $\sigma$. Data is taken from CHIME’s control run, years 16-169.
Figure 4.12: M(T,S) maps for 24 °N. From the top, the three rows indicate data representing the global longitudinal range, Indo-Pacific, and Atlantic regions respectively. From the left, the three columns represent net, northward, and southward mass transport respectively. Curved lines denote lines of constant $\sigma_t$. Data is taken from CHIME’s transient run, years 85-94.
Figure 4.13: $M(T,S)$ of different latitudes for CHIME: $34^\circ$S (A-B), $20^\circ$S (C-D), $0^\circ$N (E-F), $24^\circ$N (G-H), and $40^\circ$N (I-J). Left columns denote the entire longitudinal cross-section (Global), whilst right columns only include the Atlantic basin. Horizontal solid red and blue lines indicate values of area weighted temperatures for northward and southward moving waters respectively. Vertical dashed red and blue lines indicate the same description but for salinity. Coloured boxes for the Atlantic region correspond to unique water masses belonging to: NACW (green), AAIW (red), NADW (blue), and AABW (black).
Figure 4.14: Similar to Fig. 4.13 but for the HadCM3 model.
CHAPTER 5

A Simple Relationship between Heat Transport and the Oceanic T/S Curve

5.1 Introduction

This chapter tests new theory to describe the ratio of ocean to atmospheric heat transport at mid-latitudes simply from temperature and salinity variations, combining the analysis of ocean mass transport from the previous chapter 4 (M(T,S)) with methods of calculating meridional heat transport used in chapter 3. Motivation for this work is given below:

Estimates of oceanic and atmospheric poleward heat transport have been published over the years, yet there is generally little agreement over the order of magnitude of the contribution of the ocean and the atmosphere to the total poleward energy transport in middle latitudes. In the 70s, estimates by Vonder Haart and Oort (1973) suggested a comparable amplitude for oceanic ($H_o$) and atmospheric ($H_a$) northward heat transport at 40°N ($H_o \approx H_a \approx 2PW$). This was later confirmed by the analysis of Carissimo et al. (1985), who extended this result to extra-tropical latitudes of the Southern Hemisphere. More recently, however, Trenberth and Caron (2001) produced estimates with a much larger contribution of the atmosphere to the total poleward heat transport than the ocean in middle latitudes (for latitudes poleward of 30°).
Interestingly, the dominance of the atmosphere in middle latitudes suggested by Trenberth and Caron (2001) was hinted at in much earlier studies, for instance, in Bjerknes’ monograph on Atlantic air-sea interactions (1964), using data from Houghton (1954) and Sverdrup (1957).

On the observational side, it is clear that there are serious issues regarding uncertainties in each of the above estimates of $H_o$ and $H_a$ (e.g., Wunsch, 2005). With the advent of satellite technology, calculations of heat transport from radiation balance methods (Trenberth et al., 2001; Trenberth and Stepaniak, 2003a,b; Zhang et al., 2006) have significantly increased the accuracy of measurements, though error bars on ocean heat transport still remain at $\pm 0.3 PW$ (Trenberth and Caron, 2001).

Important theoretical work has been made in constraining heat transport by instead looking at the ratio of these two quantities instead of individually; Held (2001) derived a formula for the ratio of ocean to atmospheric heat transport that was proportional to the ratio of atmospheric gross static stability between moist and dry air parcels, though this theory was only limited to the Tropical Hadley Cell regime. Work by Czaja and Marshall (2006) analysed this ratio by decomposing heat transport into a mass stream function and energy contrast for each medium, finding a dominance of atmospheric mass transport in mid to high latitudes, and the increasing importance of the oceanic energy contrast within the Tropics.

The question of why should the atmosphere or the ocean dominate the total poleward heat transport in middle latitudes or elsewhere is still open despite recent attempts at addressing this problem; the arguments of Vallis and Farneti (2008a) illustrate that ocean heat transport can dramatically change by varying certain oceanic quantities eg. ocean basin width, diapycnal diffusivity etc, allowing for a much more varying heat transport that could have changed quite significantly within the Earth’s past. Though their scaling of atmosphere to wind driven ocean heat transport is similar to that of Held (2001), the ratio between the atmosphere to that of the buoyancy driven heat transport of the ocean (Vallis and Farneti, 2008a, Eq. 4.10) contains numerous variables to which no accepted theory exists in measuring them. “We cannot, therefore, on theoretical or a priori grounds state that the atmospheric energy transport can be expected to be larger than that of the ocean, although this may be the case in the real world” (Vallis and Farneti, 2008a). This elegantly mirrored the results of Enderton and Marshall (2009) whose calculations of heat transports on different land/ridge configurations showed the large degrees that heat transport could take.

Middle latitudes are particularly resistant to simple explanations because of the difficulty in parameterizing the heat transport by atmospheric baroclinic waves and oceanic eddies, and
because of the lack of a simple theoretical framework to predict the heat transport of the large-scale ocean circulation beyond that associated with Sverdrupian (horizontal) gyres.

In this study, we bypass these difficulties by using an earlier result on the ratio of ocean heat to freshwater transport due to Stommel and Csanady (1980). From simple arguments these authors showed that this ratio solely depends on the mean salinity of the oceans and the slope of the oceanic temperature/salinity relationship. Acknowledging that the oceanic freshwater transport is proportional to the atmospheric latent heat transport, itself a major contributor atmospheric heat transport in middle latitudes (e.g., Pierrehumbert, 2002), we obtain a simple expression for the ratio $H_o/H_a$. The latter is found typically smaller than unity, suggesting, from fairly general and robust arguments, a dominant role for the atmosphere in transporting heat poleward in middle latitudes.

The derived equations are not intended as a better method of measuring $H_o/H_a$ (since satellite data is far superior in terms of temporal and spatial coverage to work out the heat transport terms as compared to oceanic in situ data), but to show and prove a different procedure in measuring this ratio through a greatly simplified and new framework. Through this our lasting aim is to further stimulate the debate on the present theories of heat transport.

5.2 Stommel and Csanady’s analysis

We denote by $M$ and $F$ the total amount of oceanic water mass and atmospheric water mass moving poleward across a given latitude circle per unit time, respectively (Fig. 1). At mid-latitudes the transport of atmospheric freshwater through latent heat is polewards. Conservation of mass (across both ocean and atmosphere media) requires that, in the long term, the total mass of ocean moving equatorward per unit time be then $M + F$. Neglecting air-sea salt exchange, the steady state salt budget for the portion of ocean poleward of the given latitude circle reads,

$$0 = M (S + \Delta S) - (M + F) S$$  \hspace{1cm} (5.1)

in which $S$ and $S + \Delta S$ denote the (mass weighted) averaged salinity of the oceanic water masses going equatorward and poleward, respectively. This can be rearranged into

$$\frac{F}{M} = \frac{\Delta S}{S}$$  \hspace{1cm} (5.2)

A simple but powerful deduction from (5.2) is that in order to accommodate for the poleward atmospheric transport of moisture ($F > 0$), seawater moving poleward must be saltier ($\Delta S > 0$) than that moving equatorward.
Likewise, if $T$ denotes the (mass weighted) averaged temperature of these oceanic water masses $M$, and $T + \Delta T$ the (mass weighted) temperature of the oceanic water masses going poleward, the oceanic heat transport $H_o$ across the latitude circle is then,

$$H_o = c_o M (T + \Delta T) - c_o (M + F) T$$  \hspace{1cm} (5.3)$$

in which $c_o$ is the specific heat capacity of seawater.

We anticipate that $M \gg F$ (Stommel and Csanady, 1980; Nilsson and Kornich, 2008) such that using typical numbers $F = 1 \text{ Sv} \ (1 \text{ Sv} = 10^6 \text{ m}^3/\text{s})$, $S = 35 \text{ psu}$, and $\Delta S = 1 \text{ psu}$, Eq. (5.2) yields $M \approx 30 \text{ Sv}$ (as we shall see later, values of $M$ can be even much larger than this). This allows a first order approximation of (5.3) to be written as

$$H_o \approx c_o M \Delta T$$  \hspace{1cm} (5.4)$$

Using (5.4) and (5.2), and approximating the salinity of equatorward moving waters by the mean ocean salinity $S_o$, i.e., $S \approx S_o$, we readily obtain

$$\frac{H_o}{F} \approx c_o S_o \frac{\Delta T}{\Delta S}$$ \hspace{1cm} (5.5)$$

A few qualitative remarks apply before doing so. Since moisture is transported poleward in the atmosphere in midlatitudes ($F > 0$), a poleward oceanic heat transport ($H_o > 0$)
requires from (5.5) that the slope of the T/S curve be positive. This is typically the case in the North Atlantic where warm and salty waters are transported poleward in the upper layer of the column and colder and fresher waters are transported southward at depth. Conversely, an equatorward oceanic heat transport, as occurs in the South Atlantic (e.g., Bryden and Isawaki, 2001), requires from (5.5) that the T/S slope be negative. This is consistent with the presence of a salinity minimum (change of sign of the slope in the T/S curve) associated with Antarctic Intermediate Water in the Southern Ocean.

5.3 Oceanic to atmospheric heat transport ratio

5.3.1 Derivation of a simple expression for the ratio $H_o/H_a$

The energy per unit mass carried by an air parcel is, neglecting the small contribution from kinetic energy, the moist static energy $h$ (Neelin and Held, 1987),

$$h = c_p T_a + \Phi + l_v q$$ (5.6)

In (5.6), $c_p$ and $l_v$ are the specific heat capacity of dry air and enthalpy of vaporization of water, respectively, $T_a$ is air temperature, $q$ specific humidity and $\Phi$ is gravitational potential energy per unit mass. The relative contribution of dry static energy ($c_p T_a + \Phi$) and latent heat ($l_v q$) to the meridional transport of moist static energy is measured by the parameter $\gamma$,

$$\gamma \equiv \frac{c_p v T_a + v \Phi}{l_v v q}$$ (5.7)

in which $v$ is the meridional velocity of air and an overbar denotes an average along a latitude circle.

In the extra-tropical atmosphere, the dominant mechanism of energy transport across a latitude circle is through deviations from the zonal mean, i.e., eddies (e.g. Peixoto and Oort, 1992). The meridional moist static energy transport at a given pressure level is then,

$$\overline{vh} \approx \overline{v'h'} = c_p \overline{v'T_a'} + \overline{v'\Phi'} + l_v \overline{v'q'}$$ (5.8)

in which primes denote deviations from the zonal average. Since atmospheric eddies are to leading order in geostrophic balance,

$$v' \approx \frac{1}{f} \frac{\partial \Phi'}{\partial x}$$ (5.9)

$f$ denoting the Coriolis parameter and $x$ the distance in the east-west direction, there is no net transport of potential energy across a latitude circle, $\overline{v'\Phi'} \propto \overline{v' \partial \Phi'/\partial x} = 0$. Equation (5.8) and (5.7) can thus be further simplified into,

$$\overline{vh} \approx c_p \overline{v'T_a'} + l_v \overline{v'q'}$$ (5.10)
and
\[ \gamma \approx \frac{c_p v'T_a + v' \Phi}{l_v v' q'} \approx \frac{c_p v'T_a}{l_v v' q'} \] (5.11)

Integrating (5.10) in the vertical, the total atmospheric energy (or simply heat) transport across a given meridional section reads,
\[ H_a = \int \frac{v h dp}{g} \approx \int \frac{(1 + \gamma) v' q' dp}{g} \] (5.12)

Within a midlatitude baroclinic wave (storm), temperature and specific humidity perturbations are correlated, as the wave moves warm/moist air poleward and upward and dry/cold air equatorward and downward. The two terms on the r.h.s of (5.10) are thus two faces of the same process, both transporting energy poleward. Typically the contribution of temperature and moisture transports are comparable, since, using typical values of temperature and specific humidity perturbations within a storm (5-10K and 5-10g/kg, respectively), the ratio \( \gamma \approx 1 \).

At leading order in the extra-tropics, \( H_a \) can thus simply be estimated from (5.12) as
\[ H_a \approx (1 + \gamma) l_v F \] (5.13)
in which we have further used
\[ F = \int \frac{v q dp}{g} \approx \int \frac{v' q' dp}{g} \] (5.14)
to keep consistency with (5.8). Note that equation (5.13) would be a very poor approximation of the atmospheric energy transport in the Tropics, since it is primarily potential energy which is transported poleward, sensible \( (c_p T_a) \) and latent energy \( (l_v q) \) being transported equatorward by the mean meridional (Hadley) circulation.

Combination of (5.13) and (5.5) provides the relation for the ratio of oceanic to atmospheric poleward heat transport across a given midlatitude circle,
\[ \frac{H_o}{H_a} \approx \frac{c_o}{(1 + \gamma) l_v S_o} \frac{\Delta T}{\Delta S} \] (5.15)

By introducing
\[ \mu \equiv \frac{\Delta T}{\Delta S} \] (5.16)
we can rewrite (5.15) as
\[ \frac{H_o}{H_a} \approx \frac{c_o S_o}{l_v} \frac{\mu}{1 + \gamma} \] (5.17)

The ratio is seen to depend solely upon thermodynamics constants (specific heat capacity and enthalpy of vaporization), the mean ocean salinity, the slope \( \mu \) of the T/S curve drawn by the
water masses moving poleward and equatorward, and the ratio $\gamma$ of atmospheric dry static energy to latent heat transport. Only geostrophy and the hypothesis that eddies dominate atmospheric transports have been assumed. Considering the complexity of the fluid motions and diabatic processes associated with oceanic and atmospheric heat transports, relation (5.17) is a powerful and simple relation to compute their ratio.

5.3.2 Physical interpretation

The interesting aspect of the relation (5.17) is that it does not involve the ratio of oceanic to atmospheric mass transport, although the latter has been shown to be crucial in setting the partitioning of heat transport in midlatitudes (Czaja and Marshall, 2006). To see how this can be, we consider the limit in which the atmospheric heat transport is dominated by the transport of latent heat, $H_a \simeq l_v F$. In this limit, the atmospheric heat transport is achieved by transporting meridionally air masses of different moisture content ($\Delta q$), the associated mass transport $M_a$ being set by the intensity and scale of the baroclinic waves. We can thus write,

$$\frac{H_o}{H_a} \simeq \frac{H_o}{l_v F} = \frac{H_o}{M_a l_v \Delta q} = \frac{M_o e_a \Delta T}{M_a l_v \Delta q}$$

(5.18)

A comparison of this expression with (5.15) shows then that with $\gamma \to 0$,

$$\frac{\Delta S}{S_o} = \Delta q \frac{M_a}{M_o}$$

(5.19)

This equation makes it clear that the information on the ratio of mass transports is indeed implicit in the oceanic salinity contrast. This arises because, salt not being exchanged at the air-sea interface, the salinity of water moving poleward must be greater than that going equatorward in order to accommodate for the poleward atmospheric transport of water vapour $F$—Eq. (5.1). Since $F$ is proportional to $M_a$, the salinity balance in effect couples $M_o$ and $M_a$. Note that there is nothing special about salt here: a similar constraint on the ratio $M_o/M_a$ would be found for any other tracer not having surface or interior sources and sinks.

Any effects of changing the diapycnal diffusivity of the ocean $K_v$ are taken into account within the theory; to illustrate this case we use a simple thought experiment whereby $K_v$ increases, leading to an increase in $H_o$. From Bjerknes Compensation arguments, atmospheric heat transport decreases to compensate, thus the ratio of $H_o/H_a$ increases. The increase in $K_v$ leads to greater ocean mixing, resulting in decreases to both $\Delta T$ and $\Delta S$. Assuming that these two quantities change by similar amounts, the ratio of the two remains constant. The increase in ocean heat transport leads to high latitudes warming faster (Polar Amplification), resulting in a reduction to the equator to pole temperature gradient $\Delta T_e$, that drives a similar decrease
in the transport of DSE. This effect coupled with the increase in latent heat transport with a higher surface temperature leads to a reduction in $\gamma$. Thus using (5.17), the ratio increases as predicted.

### 5.4 Test of the Relationship in a Coupled Climate Model

#### 5.4.1 Heat Transport and Partitioning in HadCM3

The validity of (5.17) was tested using the Hadley Centre HadCM3 coupled climate model (see Chapter 2, section 2.2 for details). We used the years 640 - 650 from spin-up of the control run simulation to calculate a monthly mean climatology for all model calculations. We note that this is a different control climate to that used in Chapter 3.

The oceanic heat transport $H_o$ was calculated from the horizontal divergence of surface flux at the ocean-atmosphere interface (we removed ocean heat storage by subtracting the temporal and spatial mean surface flux value, $0.02 \text{W m}^{-2}$, from the incoming surface flux); we name this version of heat transport, $H_o^{\text{true}}$, where the ‘true’ superscript corresponds to the total ocean heat transport (Fig. 5.2, solid). The ocean heat transport was also computed using (5.4), to which we name, $H_o^{\text{reso}}$ (Fig. 5.2, dotted), where the ‘reso’ superscripts correspond to this method measuring the resolved advective term of heat transport (see next subsection for details on $M$ and $\Delta T$). We note that for this control run of HadCM3, only the resolved advective term of meridional velocity was available, not including “bolus” velocity terms linked to sub-grid mesoscale eddy processes. The latter was indirectly measured as the difference between $H_o^{\text{true}}$ and $H_o^{\text{reso}}$ in Fig. 5.2. Implied eddy heat transport within northern hemisphere extra-tropical latitudes are almost negligible, whereas it is comparable to $H_o^{\text{res}}$ for the southern hemisphere. The direction of parameterised eddy heat transport is consistent with ocean eddies transporting heat equatorward at low latitudes and poleward at high latitudes (e.g. Jayne and Marotzke, 2002).

Both methods of calculating ocean heat transport illustrates an antisymmetric distribution about the equator, with the northern hemisphere transporting almost twice as much heat within the tropical regions (peak magnitude of $1.7 \text{PW}$ at $15^\circ N$) as compared to its southern hemisphere counterpart. The larger northern hemisphere heat transport stems mostly from the Atlantic basin in which the Southern Atlantic Ocean advects heat northwards, opposing the other basins’ poleward heat transport, as discussed previously (section 5.2).

To compute the atmospheric heat transport $H_a$ (Fig. 5.2, black dashed) we use the same
Figure 5.2: Meridional Heat Transport computed from HadCM3 decomposed into components for the atmosphere (dashed), the ocean computed from surface flux divergence (solid), and the ocean resolved advective term (dotted).

method as Wunsch (2005) who indirectly measured the latter as the difference between total heat transport (not shown) and $H_o^{true}$. The total heat transport was calculated using the same methodology used for $H_o^{true}$, with a temporal and spatial mean flux value of $-0.06 W m^{-2}$ at the top of the atmosphere. As found for the ocean heat transport, the distribution of $H_a$ is antisymmetric about the equator, with peak magnitudes $4.6 PW$ at $37.5^o N$, and $-4.7 PW$ at $40^o S$. Within the deep tropics, the atmosphere transports very little heat energy polewards due to the almost near cancellation of latent and dry static energy within the tropical Hadley Cells. The ocean therefore is the main carrier of heat within the tropics. The atmospheric role to transport heat becomes increasingly important the further poleward, with baroclinic instabilities at mid-latitudes being the main driver. The atmosphere is therefore the main carrier of heat for extra-tropical latitudes.

5.4.2 Calculations of $M_o$, $F$, $\Delta T$, $\Delta S$, and $\mu$

Calculations of $M_o$, $F$, $\Delta T$ and $\Delta S$ were carried out using the model outputs of meridional velocity $V$, $T$ and $S$ being interpolated onto a regular fine ($\Delta Z=10 m$) vertical grid. The vertical resolution of 10m thickness was chosen to accurately represent the layers closest to the ocean surface. Mass conservation was enforced across a latitude circle by subtracting from $v$ its average value over the section. For each latitude we plotted meridional mass transport as a function of temperature and salinity bins, $M(T,S)$, using

$$M(T,S) = \int_0^{2\pi} \int_0^\theta \int_{-T}^{T+\delta T} \int_{S-\delta S}^{S+\delta S} \rho v dz dT dS'$$

(5.20)
where H represents ocean depth, and temperature and salinity resolution is of the order \( \delta T = 0.2^\circ C \) and \( \delta S = 0.1psu \) respectively. Fig. 5.3 shows the resulting mass transport at \( 40^\circ N \). One distinguishes clearly between the warm and salty water masses found within the North Atlantic \( (S > 35psu) \), and the relatively cool and fresher water masses within the North Pacific \( (S < 35psu) \). To compute poleward flowing mass weighted values of temperature \( T_p \) and salinity \( S_p \), we find the ‘centre of mass’ point for poleward flowing waters on the \( M(T, S) \) plot, whose coordinates correspond to \( T_p \) and \( S_p \). As an example, for Fig. 5.3 the centre of mass point for poleward flowing waters is centered at \( 34.7psu \) and \( 6.3^\circ C \) (intersection between horizontal black solid and vertical black dashed line), therefore at \( 40^\circ N \), \( T_p = 6.3^\circ C \) and \( T_p = 34.7psu \). To compute equatorward flowing mass weighted values, \( T_e \) and \( S_e \), we follow the same methodology (white lines in Fig. 5.3), but instead compute the centre of mass point for all equatorward flowing values of \( M(T, S) \). From these values we readily obtain \( (\Delta T, \Delta S) = (T_p, S_p) - (T_e, S_e) \).

The latitudinal distributions of these variables can be seen in Fig. 5.4, with \( \Delta S \) and \( F \) being multiplied by a factor of 10 and 1000 respectively to magnify detail; black (grey) lines represents the northern (southern) hemisphere. The latitudinal axis for the southern hemisphere is flipped to have a shared axis with northern hemisphere values.

Northern Hemisphere averaged \( \Delta T \) is on the order of \( 2^\circ C \) for the extra-tropics (Fig. 5.4a, black, solid). Although not shown the Atlantic basin has the largest temperature contrast of \( \Delta T \approx 4^\circ C \), with the Indo-Pacific basin having a combined \( \Delta T \approx 1^\circ C \). Similarly an average

![Figure 5.3: Plot of Mass Transport \( M \) as a function of salinity (x-axis) and temperature (y-axis). Mass weighted values of temperature (salinity) for meridional moving waters are represented by the solid (dashed) lines. Dark (light) shading refers to poleward (equatorward) flowing waters. Values of mass transport are divided by a constant density \( \rho \) to have the same units of volume flux measured in Sv.](image-url)
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Figure 5.4: Distribution of $\Delta T$, $\Delta S$, meridional mass transport $M$, and freshwater transport $F$ in the atmosphere. Northern (Southern) hemisphere values are represented by black (grey) lines. $\Delta S$ (a) is multiplied by a factor of 10 to highlight the salinity features, with the southern hemispheric freshwater transport being multiplied by a factor of 1000 for similar reasons. The latitudinal axis for the southern hemisphere plots are reflected such that they overlap that for the northern hemisphere.

$\Delta S \approx 0.2\text{psu}$ (Fig. 5.4a, black, dashed) stems largely from the Atlantic basin with values of $\Delta S \approx 0.3\text{psu}$ and $\Delta S \approx 0.1\text{psu}$ for the Indo-Pacific region. This is consistent with the fact that the Atlantic basin is saltier and warmer than all others. For the Southern Hemisphere (grey curves), both quantities are reduced especially $\Delta T$, with an average $\Delta T < 0.1$, and $\Delta S$ a factor 2 smaller than in the Northern Hemisphere. Southward of $64^\circ S$ salinity increases with depth to a maximum of $\approx 35\text{psu}$ at 4500m. At these latitudes, the equatorward flow of waters originating within the Ross and Weddell Sea’s extend almost to the bottom of the sea floor. These deep equatorial flowing waters pick up the high salinity values associated with depth, leading to a reversal in sign of $\Delta S$ at $64^\circ S$.

Mass transports (here expressed in $Sv$ by dividing by a constant density $\rho$) for the northern hemisphere ranges between $200Sv$ at mid-latitudes to $20Sv$ at higher latitudes (Fig. 5.4b, black, solid). For the southern hemisphere, a peak in meridional mass transport of $700Sv$ is found at $57^\circ S$. It should be emphasised that $M$ represents the summation over all northward or all southward velocity cross sectional areas; this is to be contrasted with the standard Eulerian mean calculation which computes the net values of northward minus southward mass transport at every depth layer. The increase in meridional mass transport for the Southern Hemisphere region corresponds to there being a larger fraction of ocean area (this be equal to 100% ocean area for the Southern Ocean region). The Northern Hemisphere ocean heat transport, when compared to its Southern counterpart, differs therefore by its large temperature contrast and
small mass transport.

The atmospheric freshwater fluxes $F/Sv$ (Fig. 5.4b, dashed) were computed using (5.2) ie. $F = M_o \Delta S/S_o$. Both hemispheres have similar magnitudes near the Tropical boundaries, and display an almost complete reduction near polar latitudes. This decrease is particularly pronounced in the southern hemisphere, with cold Antarctic temperatures decreasing the saturation specific humidity of air. The Southern Hemisphere does however transport more than twice as much freshwater than the Northern Hemisphere poleward of 45°S, possibly due to the greater presence of ocean surface and tropical rainforest regions (i.e. the Amazon basin) for the former hemisphere.

![Figure 5.5: Distribution of the T/S slope $\mu$ (a), and the model observed values for the ratio of dry static energy to latent heat transport in the atmosphere, $\gamma$ (b). Northern (Southern) hemisphere values are shown in black (grey), with true $\gamma_F$ (solid) computed from freshwater transports, and predicted $\gamma_T$ (dashed) computed from zonally averaged surface temperatures.](image)

The ratio $\mu \equiv \Delta T/\Delta S$ (Fig. 5.5a, black) is greater in the northern than in the southern hemisphere, this being mainly due to the large temperature contrast $\Delta T$, originating from the meridional overturning circulation found in the North Atlantic. The reduced values of $\mu$ for the southern hemisphere reflects the greater reduction in $\Delta T$ compared to $\Delta S$. The sudden drop in $\mu$ at 64°S corresponds to where $\Delta S$ changes sign. It is interesting to note that even though the observed ratio of $H_o/H_a$ is similar for both hemispheres, values of $\mu$ in each hemisphere vary by almost an order of magnitude at some latitudes. According to (5.17) the large fluctuations for the latter term must be partially balanced by changes in $\gamma$ to balance out the heat transport ratio $H_o/H_a$ (we come back to this issues in the discussion section).
5.4.3 Test of the Formula

Fig. 5.6 compares the model distribution of $H_o/H_a$ (solid) with the predicted ratio (dashed) computed using (5.17) and using $\gamma = (H_a - l_v F)/l_v F$, where $F$ is the freshwater transport shown in Fig. 5.5b; this expression will be referred to as $\gamma_F$. Northern Hemisphere $H_o/H_a$ values (black) show a close agreement between predicted (dashed) and observed (solid) quantities. The minimum at $40^\circ N$ coincides with the locations of the major western boundary currents e.g. the Gulf Stream and Kuroshio. These fast moving currents are warmer than their surroundings and therefore provide a source of heat for the atmosphere. These help fuel the mid-latitude storms that are blown eastward, while at the same time reducing the heat transport by the ocean. In a more “gyral” explanation, the separated western boundary currents transports most of its heat zonally, thus having a vanishing meridional component of velocity. Both explanations help explain the decrease in $H_o$ and therefore a subsequent decrease in (5.17).

The ratio of ocean to atmosphere heat transport in HadCM3 is smaller in the Southern than in the Northern Hemisphere, being due to the smaller contribution of $H_o$ at these latitudes by almost a factor of 2, whilst the atmospheric component increases to compensate, leaving an almost perfect anti-symmetric distribution of total heat transport. Polewards of $40^\circ S$ predicted values of (5.17) do not fit well with the true model values, owing to the large parameterized eddy component of heat transport within the southern hemisphere (Fig. 5.2, section 4a) which was not taken into account in the calculations of $M_o$, $\Delta T$ and $\Delta S$. 

Figure 5.6: The ratio $H_o/H_a$ in HadCM (solid) and predicted from (5.17) using $\gamma_F$ (dashed). Plots in black (grey) represent the Northern (Southern) Hemisphere. The dotted lines corresponds to an additional calculation in which $\gamma = \gamma_T$, is predicted rather than calculated in the model.
5.4.4 Constraining the Theory even further

Equation (5.17) provides a good approximation to $H_o/H_a$ using $\Delta T$, $\Delta S$ and $\gamma$. In order to simplify (5.17) even further by using solely temperature and salinity as variables, a new expression for $\gamma$ that is solely dependent on surface temperature $T_s$ was developed. By setting a constant relative humidity $RH$, one can scale fluctuations in specific humidity $q'$ to fluctuations in lower tropospheric air temperatures, analogous to warm (cold) air carrying more (less) moisture. Through invoking the Clausius-Clapeyron relationship and approximating lower tropospheric air temperature as similar to $T_s$, we compute a new expression for $\gamma$, hereafter denoted by $\gamma_{T_s}$

$$\gamma_{T_s} = \frac{c_p R}{L_v RH q_{sat}(T_s)} \frac{T_s^2}{T_s}$$

(5.21)

where $R = 287 \text{ J kg}^{-1} \text{ K}^{-1}$ is the gas constant, $RH = 0.8$ is constant relative humidity, $q_{sat}$ is the saturation specific humidity, and $T_s$ is the zonal average surface temperature – the full derivation of this equation is given in the Appendix, section B. The resulting heat transport ratio using this new function of $\gamma_{T_s}$ is shown in Fig. 5.6 (dotted), displaying a comparable skill as when using $\gamma_F$.

5.5 Application of the Theory

5.5.1 Use of Oceanic Observations

Testing the relation (5.17) in observations might at first sight seem daunting, since it requires a knowledge of global, three dimensional meridional mass transports in order to estimate $\Delta T$ and $\Delta S$. We propose here an alternative application of (5.17) to the Southern ocean using the assumption that in some latitude band, the ocean heat transport is dominated by its eddy component. Such assumption is reasonable near 45°S as recent high resolution ocean models have suggested (e.g., Jayne and Marotzke, 2002; Meijers et al., 2007).

Under the assumption that it is meso-scale eddies, rather than quasi-steady gyres or overturning cells, which are the major contributors to the ocean heat transport near 45°S, the ratio $\Delta T/\Delta S$ associated with oceanic eddies can be estimated from lateral gradients in temperature and salinity. Let us denote by $l$ the typical excursion of water parcels within an eddy in the (meridional) $y$ direction. The temperature difference $\Delta T_{eddy}$ between equatorward and poleward flowing parcels is then

$$\Delta T_{eddy} = 2l \times \frac{\partial T}{\partial y}$$

(5.22)
Application of a similar reasoning to salinity contrasts $\Delta S_{\text{eddy}}$ yields a simple relation for the ratio $\Delta T/\Delta S$ appearing in (5.15),

$$\frac{\Delta T}{\Delta S} \approx \frac{\Delta T_{\text{eddy}}}{\Delta S_{\text{eddy}}} = \frac{\partial T}{\partial y} \frac{\partial S}{\partial y}$$

(5.23)

Figure 5.7: Meridional salinity (in psu) and potential temperature (in degree Celsius) contrasts across $45^\circ S$ as estimated from WOCE section’s S05. The dashed line indicates the $8K/\text{psu}$ line.

To estimate the oceanic to atmospheric heat transport ratio at $45^\circ S$, we thus need to estimate the ratio of temperature to salinity gradients over the layer in which eddy-induced lateral displacements are significant (0-1000m, e.g., Wunsch, 1999). In Fig. 5.7, we present a scatterplot of observed meridional temperature/salinity changes across the Sub-Antarctic Front at $45^\circ S/120^\circ E$, estimated as $\Delta T_{\text{eddy}} \approx T(46^\circ S) - T(45^\circ S)$ and $\Delta S_{\text{eddy}} \approx S(46^\circ S) - S(45^\circ S)$, using the data taken during the World Ocean Circulation Experiment (WOCE, see White, 1994) with a roughly $0.5^\circ$ resolution in latitude. The scatterplot is seen to be composed of three well separated regions, indicated by the black circles (0-600db), grey circles (600-700db) and black stars (700-1000db). Within the upper layer (black circles, including the mixed layer, seasonal thermocline and a well-defined layer of sub Antarctic mode water), a sharp linear relation is found with a value $\mu \equiv \Delta T/\Delta S = 8K/\text{psu}$, over a broad range of temperature (0 to 2.5K) and salinity (0 to 0.3 psu) contrasts. The lower layer (black stars) shows a similar range of salinity contrasts but twice stronger temperature contrasts.

We further need an estimate of $\gamma$ in order to predict the ratio $H_o/H_a$ from (5.17). This is obtained from atmospheric reanalyses data with the results that, at $45^\circ S$, $\gamma \simeq 2PW/2.6PW = 0.77$ (Pierrehumbert, 2002; by inspection of his Fig. 1a). Application of (5.17) at $45^\circ S$ thus yields,

$$\frac{H_o}{H_a} \approx \frac{4000 \times 35}{2.5 \times 10^6} \times \frac{8}{1 + 0.77} \approx 0.25$$

(5.24)
which is in agreement with a dominance of the atmospheric contribution to the total poleward heat transport in the Southern Hemisphere (e.g., Trenberth and Caron, 2002; Czaja and Marshall, 2006).

5.5.2 Warm Paleo-Climates

For an extreme in warmer climates where latent heat transport dominates DSE, the quantity $\gamma < 1$. The ratio of ocean to atmospheric heat transport can be seen as proportional to the T/S curve (as $\mu >> \gamma$), and as such, only the knowledge of the temperature and salinity (or any other conserving tracer) gradients of the ocean would be necessary in determining the ratio of ocean to atmospheric heat transports. This method may be of use to paleoclimatologists eg. the warm Eocene period, where oceanic proxy measurements of such quantities could be used to reconstruct historical values of $H_o/H_a$, offering a further tool to glimpse into our past climate. An optimum location to use this would be the Southern Ocean region (see previous subsection) whereby the dominance of eddy activity allows only the lateral gradients of salinity and temperature needed to calculate the T/S slope.

This method however works both ways; since presently it would be easier to measure heat transport from satellite data rather than the T/S curve indirectly, so could the greater accurate measurements of $H_o/H_a$ be used to indirectly monitor the status of the ocean's T/S curve.

5.6 Discussion

Besides its domain of validity being restricted to midlatitudes, there are three main limitations of the formula (5.17) that we wish to discuss. First, it is based on the assumption that the coupled ocean-atmosphere freshwater balance has reached equilibrium –see eq. (5.1). “Hosing experiments” in which the North Atlantic overturning circulation is artificially reduced (e.g., Stouffer et al., 2006) show that this might only be achieved on timescales of centuries. Our comparison of predicted and true ratio $H_o/H_a$ for the northern hemispheric extra-tropical latitudes indicates discrepancies on the order of 10 – 20%, depending on the choice of $\gamma$ (Fig. 5.6) used in (5.17). These errors originate from the fact that the hydrological cycle has not completely reached equilibrium yet, i.e. that in practice that there are differences between the indirect calculation of atmospheric freshwater transport $F$ using (5.2), and the direct calculation using evaporation minus precipitation (by inspection of Pardaens et al. 2003, Fig. 3).

Second, although the framework used to derive (5.17) allows ocean eddies to be included,
it does so only partially. A rough observational estimate of the advective effects of eddies was proposed in section 5 (and, in an ocean model framework, the “bolus” velocities associated with eddy parameterization could be used explicitly to calculate $M$ and $F$), but it is impossible to include the diffusive effect of the eddies (e.g., Plumb and Mahlman, 1987) in a mass transport framework.

Thirdly, a significant drawback of the theory is that Eq. (5.17) is only viable where the latitude of interest is sufficiently warm such that no ice cover is present within the surrounding region. Ice cover does not allow for communication of freshwater between the ocean and atmosphere and therefore salinity contrasts cannot be achieved. In the extreme case of a ‘Snowball Earth’ scenario where the planet is completely covered in ice, the complete isolation of freshwater exchange at the surface would result in a homogenization of salinity within all oceans, thus $\Delta S \to 0$, rendering Eq. (5.17) useless. A second consequence of this cold climate scenario would be that Latent Heat $\to 0$ due to the reducing ability of the atmosphere to hold moisture, leading to $\gamma \to \infty$. Although surprisingly this leads to $H_o/H_a \to 0$ whereby all heat transport in carried by the atmosphere (a true statement for this climate), the surface ice conditions again renders the measurements of $\Delta T$ and $\Delta S$ meaningless. The formulation of Eq. (5.17) is therefore limited to ‘warm’ climate scenarios only.

Inspection of the variables $\mu$ and $\gamma$ (section 4b) revealed that their magnitudes are both larger for the Northern than for the Southern Hemisphere. Although the observed model ratio $H_o/H_a$ is approximately twice as large within the Northern Hemisphere, $\mu$ and $\gamma$ vary by almost a factor of 8 and 3 respectively when comparing hemispheres (e.g. at $40^\circ N, S$), suggesting a form of compensation between the two variables. A naive view to explain this compensation could be as follows: we rewrite our expression for the ratio of dry static energy to latent heat transport in the atmosphere $\gamma$, as $\gamma = H_a - l_v M \Delta S / l_v M \Delta T$, where dry static energy is computed as the residual between atmospheric and latent heat transport, and freshwater transport $F$ is rewritten using (5.2). Since $\gamma$ and the T/S slope $\mu = \Delta T / \Delta S$ thus appear inversely proportional to $\Delta S$, this latter quantity would be a suitable candidate for the coupling between the two terms (given all other quantities remain unchanged). However if we compare $\Delta S$ for the Northern and Southern Hemispheres we find that $\Delta S$ is larger for the former region, giving smaller values for $\mu$ and $\gamma$; the opposite to what we would expect from a simple $\Delta S$ coupling. The actual reduction in $\mu$ for the Southern Hemisphere is thus primarily controlled by the reduction in $\Delta T$. To explain the reduction in $\gamma$ we showed from
our previous expression of $\gamma$ above, that $\gamma$ is both inversely proportional to $\Delta S$ and meridional mass transport $M$. By inspection, differences between each hemisphere’s values for $\Delta S$ and $M$ are approximately 1/3 smaller and 7 times larger respectively, in the Southern than in the Northern Hemisphere. The reduction in $\gamma$ for the Southern Hemisphere is therefore primarily controlled by the increased mass transport $M$ being greater than the reduction in $\Delta S$ within the region.

5.7 Summary and Conclusion

Building upon the analysis of Stommel and Csanady (1980), a simple relation between the ratio of poleward heat transport by the ocean ($H_o$) and atmosphere ($H_a$) and the oceanic T/S curve was derived –eq. (5.17). This relation was tested with much success against the control simulation of the coupled climate model HadCM3 (albeit better represented within the Northern Hemisphere than for the Southern), and hydrographic observations at 45$^\circ$S. The inclusion of ‘bolus’ velocities are seen as particularly important to the accuracy of the theory within the latter region where eddy activity has a large influence.

The ratio $H_o/H_a$ was shown to depend solely on two parameters: the ratio $\gamma$ of dry static energy to latent heat transport, and the slope $\mu$ of the T/S curve drawn by equatorward and poleward moving water masses. Within a much warmer climate where latent heat greatly exceeds the transport of dry static energy within the atmosphere it was proved (though not shown) that the ratio $H_o/H_a$ could be simplified further to being proportional solely to the T/S curve.

Furthermore using simple assumptions of geostrophy and the dominance of eddy transport within the mid-latitude atmosphere, we derived a new expression for $\gamma$ that is seen to be solely dependant on the zonal mean surface air temperature.

Different oceanic and atmospheric processes (eddies, zonally averaged circulations) operate at different values of the parameters $\mu$ and $\gamma$. For example, it was shown that Southern ocean eddies are associated, at 45$^\circ$S, with values of $\mu \simeq 8 K/psu$, larger than those of the large scale circulation ($\simeq 1 K/psu$, see Fig. 5.5, grey). Comparison of both hemispheres indicated that these two terms show a partial compensation in (5.17), helping to keep the ratio $H_o/H_a$ almost symmetric about the equator. More work is warranted to explain this compensation.

The relation (5.17) offers a new perspective on the mechanisms of poleward heat transport in the climate system. Given the complexities of present theory surrounding the measurements of heat transport, particularly at mid-latitudes, this relation provides a surprisingly simplified
alternative, albeit most likely to be of first order approximation. How the formulae will fair within warmer climate scenarios and of different climate models will be the subject of future research.
CHAPTER 6

Meridional Heat and Freshwater Transport in an Increasing $CO_2$ Environment

This chapter deals with how the distribution of meridional heat and freshwater transport changes in an increasing $CO_2$ environment. Details of the scenario and the specific time mean periods used are explained in section 2.5.2 and section 2.5.3.

To understand changes in both heat and freshwater transport, we first discuss the major changes for global diagnostic variables; understanding these changes will illicit better insight into the mechanisms of ‘global warming’ when analysing for example, trends in $M(T, S)$.

6.1 Global Mean Diagnostics in a Global Warming Scenario

6.1.1 Surface Air Temperature

Figure 6.1 shows the change in air temperature at 1.5 m above ground for different transient climate states. For the majority of the Earth’s surface there is a gradual increase in near surface air temperature with increasing $CO_2$. Land masses tend to warm much faster than for the ocean, which advects its heat below, increasing the internal abyssal temperature. Figure 6.2 illustrates the zonal mean temperature anomalies for the previous plot, with dashed curves
giving the area weighted global mean surface temperature change. Northern Hemisphere polar regions warm at a faster rate compared to regions near the equator, mirroring the effects of Polar Amplification. For a doubling of CO$_2$ (blue curves) changes to near surface temperature are approximately 1.5$^\circ$C near the equator, increasing to maximum of 7.4$^\circ$C at high northern latitudes - an increase 5 times as large. Similarly, global mean near surface temperatures change by roughly 1.8$^\circ$C, or at a rate of roughly 0.25$^\circ$ per decade (assuming 70 years to reach a doubling of CO$_2$). This is comparable, albeit slightly higher, to estimates quoted in the IPCC AR4 report of 0.2$^\circ$ per decade (Solomon et al., 2007).

Polar Amplification is not seen in the Southern Hemisphere. Near 60$^\circ$S the model observes a slight cooling in near surface temperatures where the upwelling of abyssal water in the Southern Ocean from Ekman Suction overrides any regional warming trends. Northwards of about 55$^\circ$S the Southern Ocean warms. This dipole pattern was also observed in work by Fyfe et al. (2007). The latter authors reasoned that Southern Ocean warming was not due solely on anthropogenical forcing but also due to the strengthening and poleward shift of the Westerly Winds, which enhanced purely CO$_2$-induced warming. They found that changes to the wind stress curl associated with strengthening zonal winds enhanced purely CO$_2$-induced warming around 45$^\circ$C through increased downwelling of warm surface waters, and at higher latitudes, reduced warming through the upwelling of cold waters. As a comparison, Gregory et al. (2005) found similar cooling trends over the Southern Ocean from analysis of the current CMIP models of the time, however this was assumed to be the consequence of an enhanced hydrological cycle that increased high latitude precipitation, thus cooling the surface.

From work in section 3.6 we expect major feedbacks in the atmosphere with a possible strengthening of the SH Hadley Cell as a result. Other regions of relative insensitivity to SST changes include the western coast of South America where again, coastal upwelling of cold water negates any SST changes.

### 6.1.2 Sea Surface Salinity

Although the majority of changes in near surface temperature are positive, the relationship between temperature and salinity is not straight forward, with trends in SSS (Fig. 6.3) having its own unique characteristics, devoid of clear mechanisms associated with Polar Amplification. The major trends associated with SSS include a gradual salinification of the entire Atlantic basin region and parts of the Arctic Ocean, and a corresponding freshening of the Indo-Pacific basin and Southern Ocean regions. Changes in SSS happen at a faster rate in the Atlantic
Figure 6.1: Near surface air temperature anomalies for the different transient CO$_2$ concentrations of CHIME.

Figure 6.2: Zonal mean values of Fig. 6.1. Global mean values are represented by the dashed lines.

rather than the Indo-Pacific region, with the latter region observing almost no change in SSS at 1.4 $\times$ CO$_2$ (Fig. 6.3a). The slight increases in SSS freshening start to emerge near South-East Asia and off the western coast of North America, suggesting an enhancement of land precipitation with these regions that return to the ocean as run-off. At 3 and 4 $\times$ CO$_2$ concentration, large areas of freshening occur off the coast of Greenland and the Equatorial Pacific. Due to a similar magnitude change in SSS for the Atlantic basin at similar equatorial latitudes, inter-basin salinity dipole pattern would suggest a net freshwater transport from the Atlantic into the Indo-Pacific basin by the Trade Winds (see section 3.8.2 for the complete
mechanism description.

![Figure 6.3](image)

**Figure 6.3:** Similar to Fig. 6.1 but for sea surface salinity.

### 6.1.3 Evaporation minus Precipitation

The anomalous changes to surface evaporation minus precipitation rates are given in Fig. 6.5. In the latitudinal direction the main trends include an increase in evaporation rates for the Tropics, with a corresponding increase in precipitation over the Extra-Tropics. Closer to subtropical and equatorial latitudes, a distinct pattern emerges of high net rates of evaporation south of the equator, with a corresponding net precipitation just northwards. The above pattern implies an anomalous large net transport of atmospheric freshwater crossing from the south to the northern hemisphere, particularly for the Pacific. We assume that this feature is associated with a northward shift of the ITCZ region due to the relative strengthening of the SH Hadley Cell compared with its northern counterpart. To test this, figure 6.4 plots the distribution of zonal mean precipitation for the global mean area and for the Equatorial Eastern Pacific (EEP) region bounded by 200° – 260°E (see section 6.3.1 for more details about the latter region). To simplify the detail, only the control (black and green lines) and 4 × TRANS (blue and red lines) scenarios are included to better visualise the maximum changes in the distribution of precipitation.

Analysis of the global mean position of the ITCZ region for both control and 4 × CO₂ scenario reveal no change, with both being centred at 7.5°N (black and blue lines). It is only when we average over the EEP region that a shift in ITCZ is visible, moving by 2.5°
Figure 6.4: Zonal mean distribution of precipitation computed from the global longitudinal integral (black and blue), and for the region bounded by 200 – 260°E. Only the control and 4 × TRANS scenarios are shown. Arrows represent the latitudinal position of the ITCZ for each scenario.

Shifts in the ITCZ are thus only visible over regional scales. For both 4 × TRANS scenarios, precipitation input across 7.5°N increases relative to the control run, whilst the secondary maximum near 7.5°S decreases, inferring a northward migration of freshwater through the equatorial region.

Figure 6.5: Similar to Fig. 6.1 but for evaporation minus precipitation.
6.1.4 Clausius-Clapeyron Scaling

Changes in the global mean column integrated specific humidity were computed in the atmosphere for each transient scenario and were found to increase at a rate proportional to that of Clausius-Clapeyron (C-C) - Fig. 6.6. Each data point represents a transient climate, with increasing values being proportional to the $CO_2$ concentration. CHIME’s ability to follow C-C scaling makes it consistent with the responses of IPCC AR4 ensemble models tested by Held and Soden (2006). Although the latter researchers were taking the mean difference between the years 2080-2099 and 2000-2019 for an A1B climate scenario, this is still similar to changes for a doubling of pre-industrial $CO_2$ values that are included in this report.

Percentage increases to global mean precipitation (blue stars) are at an expected lower rate to that of CC (Trenberth, 1998; Allen and Ingrin, 2002; Held and Soden, 2006). For the 4 transient scenarios, this averages out to a mean rate of approximately $1\% K^{-1}$. The empirical ensemble median value derived by Held and Soden (2006) was calculated at $1.7\% K^{-1}$, putting CHIME’s sensitivity to its hydrological cycle in the lower end of the model spectrum.

![Figure 6.6: Percentage increase of global mean column integrated specific humidity (black stars) and global mean precipitation (blue stars) as a function of changes in global mean near surface temperatures for the 4 transient scenarios. Black dashed line represents the rate of Clausius-Clapeyron, whilst the blue dashed line is an empirical derived scaling of $1\%$ change $K^{-1}$.

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6.2 Meridional Heat Transport

6.2.1 Total, Ocean, and Atmospheric Heat Components

Total poleward heat transport, and its components due to the ocean and atmosphere are computed using the same method used for CHIME in section 3.6. The distribution of each component for the 1 control and 4 transient scenarios are shown in Fig. 6.7. Total poleward heat transports are represented by solid lines, whilst the ocean and atmosphere are represented by ‘dot-dash’ and dashed curves respectively; different colours signify different CO\textsubscript{2} concentrations. To aid in explaining the changes to Fig. 6.7, the differences between transient and control scenarios are shown in Fig. 6.8. Each sub-figure represents a separate anomalous transient scenario, with changes in the total, atmosphere and ocean heat transport characterised by blue, red and green curves respectively; the error bars show a ± 1 standard deviation of the latitudinal dependent time series.

Anomalous changes in total poleward heat transport (blue curves) have a maximum of approximately 0.15 PW for 2×TRANS to roughly 0.4 PW for 4×TRANS, with trends indicating an overall increase in total poleward heat transport in both hemispheres, particularly at extra-tropical latitudes. The maximum standard deviations are of the order of ±0.13PW calculated near the equatorial regions. Given this variability, the enhancement of total poleward heat transport becomes statistically visible only for CO\textsubscript{2} concentrations greater and equal to a doubling of the control value.

For the ocean, anomalous changes in poleward heat transport reveal an overall weakening trend that is strongest for extra-tropical latitudes. The spread in model data for the transient scenarios calculate an error maximum of about ±0.6 PW for values of \(H_o\) that is considerably larger than that found for \(H_T\). Due to this large standard deviation, any changes in \(\Delta H_o\) seen across all transient scenarios is obscured by the error bars, with only slight statistically significant changes southward of 40°S for TRANS 3 and 4. If we become more generous with our statistical analysis, the latitudes between 40°N-0°N also reveal a weakening of ocean heat transport for the aforementioned transient scenarios.

Changes in the atmospheric poleward heat transport roughly follow similar patterns to that observed for total poleward heat transport, with increasing trends in both hemispheres. Since \(H_a\) was calculated as a residual between the total and ocean heat transport components, values of atmospheric heat transport share similar model standard deviations to that of the latter. Just like with the ocean, major changes to atmospheric heat transport become statistically
significant for model scenarios greater and equal to $3 \times TRANS$, with clear increases over the mid-latitudes in both hemispheres. These increases extend equatorward up to $20^\circ S$ for $4 \times TRANS$. Although $\Delta H_a$ changes in a similar fashion to $\Delta H_T$, the former changes are slightly larger. The picture that arises is that of changes to the ocean and atmospheric heat transports that are roughly equal in magnitude and opposite in direction, particularly for extra-tropical latitudes, mirroring the dynamics of Bjerknes Compensation (Bjerknes, 1964).

![Figure 6.7: Meridional heat transport in CHIME for the ocean (dot-dashed), atmosphere (dashed), and total (solid) heat components. Control scenario values are represented by black curves, whilst individual transient scenarios are shown by non-black colours: $1.4 \times CO_2$ (green), $2.0 \times CO_2$ (blue), $3.0 \times CO_2$ (cyan), and $4 \times CO_2$ (red).](image)

Significant Bjerknes Compensation has been observed in works such as Shaffrey and Sutton (2006), and Van der Swaluw et al. (2007). These aforementioned authors diagnosed the changes in HadCM3’s poleward heat transports (CHIME uses identical HadCM3 atmospheric model) in a multi-centennial preindustrial control run. They found the greatest anti-correlation between changes in atmosphere and ocean heat transports roughly at high northern hemisphere latitudes, with the correlation reaching a maximum at decadal time scales. In their analysis, changes to the atmosphere’s heat transport lagged behind those by the ocean, with fluctuations in the former due to variability in the Atlantic MOC strength. Since results for this chapter uses time mean data, no lag correlation can be established between the ocean and the atmosphere.

A crude approximation to calculate the amount of Bjerknes Compensation occurring in our data is used for all 4 transient scenarios. We define the Bjerknes Compensation variable, $C_{bk}$ as the ratio of changes in the ocean heat transport, to that for that atmosphere; that is, $C_{bk} = \Delta H_o/\Delta H_a$. Since changes are of equal and opposite magnitude, $C_{bk}$ is always negative.
Figure 6.8: Anomalous differences in meridional heat transport for the total (blue), atmosphere (red), and ocean (blue) heat components for the 4 transient scenarios. Error bars signify a ± 1 s.d. for the latitudinal dependent time series.
and as close to unity as possible. The plot of $C_{bk}$ is shown in Fig. 6.9; NH values are given by solid curves, whilst values for the SH are reflected onto the same axis, being represented by dashed curves. The latitudinal range of $C_{bk}$ is substantially larger for the SH with values for $3$ and $4 \times TRANS$ for this region extending down to almost $20^\circ S$. This is not shown, though the value of $C_{bk}$ remains relatively stable throughout this hidden region at $-0.4$. Trends in the amount of Bjerknes Compensation with increasing $CO_2$ is somewhat unclear, with control scenario values being consistently larger in both hemispheres than values for the transient scenarios. Neglecting the control scenario however, there is an increasing trend with $CO_2$ towards a state where $C_{bk}$ converges to unity. Surprisingly we find a larger amount of Bjerknes Compensation within the extra-tropical SH rather than for the NH, the opposite to what was found by Shaffrey and Sutton (2006) and Van der Swaluw et al. (2007). Although the main difference between this experiment and the mentioned authors are in the choice of transient and control climate states, this cannot be a main reason for the different hemisphere responses since CHIME exhibits larger values of $C_{bk}$ in the SH even in its control climate state.

Figure 6.9: The ratio of changes between ocean to atmospheric heat transport between transient and control climate scenarios.

To analyse the origins of changes in heat transport, we recall that $H_o$ is proportional to the zonally integrated meridional mass transport $M_o$, and the temperature difference between poleward and equatorward moving waters $\Delta T$. Both distributions of these quantities for all climate scenarios and their anomalous differences from the control climate are shown in Fig. 6.10 and Fig. 6.11 respectively. The reversal in sign of $\Delta T$ about the equator comes about due to it be calculated as the difference in temperature between northward and southward moving
water masses. In the NH this value is positive since warm waters flow above cold waters. However in the SH, southward moving water masses near the surface flow above colder waters moving northwards. The reversal in sign is thus a result of calculation convention. As a comparison, \( \Delta T \) would be all positive in value if the convention of poleward and equatorward moving water masses was used instead. The distribution of \( \Delta T \) has values that are more than double in size for the NH compared with the SH. The large inter-hemisphere differences arise from the large MOC of the Atlantic basin that transport warm surface water polewards and cold water equatorwards. In contrast to the Southern Hemisphere, the Southern Ocean has high mixing properties of all the major water masses, tending to homogenise temperature properties within the water column and thus reducing . Changes to are on the order of \( \pm 0.2 ^\circ C \).

For the statistically significant regions relating to Bjerknes Compensation described in Fig. 6.8, that is latitudes, 20\(^\circ\)S-60\(^\circ\)S and 40\(^\circ\)N-60\(^\circ\)N, there is a decreasing trend of the temperature gradient. The latitudinal temperature distributions of both poleward and equatorward moving waters (not shown) reveal a clear warming trend. The decrease in the temperature difference would therefore suggest that equatorward moving water masses are warming at a faster rate that their poleward moving counterparts, further suggesting that the ocean interior is warming. The large differences in northward of 55\(^\circ\)C is consistent with the curvilinear ocean grid of CHIME creating spurious error when calculating ocean properties through zonally integrated cross-sections. We do not dwell too much on the large anomalous differences to the temperature gradient at these latitudes since the cosine of latitude renders any calculation of heat transport very small.

Due to mass conservation the amount of rate of poleward moving water through a latitudinal boundary is the same for the equatorward moving return flow. Fig. 6.11a thus shows only the poleward component of \( M \). Comparing mirror latitudinal points, the SH has considerably larger mass transports than its NH counterpart.

Anomalous changes to ocean mass transport northwards of 40\(^\circ\)N reveal a decreases of approximately 10% for all scenarios. No clear trend between reduction in \( M_o \) and \( CO_2 \) concentration can be established, with larger decreases in \( M_o \) seen for scenarios with lower values of \( CO_2 \). For latitudes southward of 40\(^\circ\)S there is an enhancing trend of \( M_o \) across all scenarios particularly for Southern Ocean region. The enhancement meridional temperature gradient across the SH (see Fig. 6.2) would lead to an enhancement of the Westerly Winds driving an increased equatorward Ekman transport, explaining the enhanced \( M_o \) across the region. The fact that ocean heat transport decreases in the SH whilst mass transport increases, suggests
Figure 6.10: a). Latitudinal distribution of the temperature gradient between poleward and equatorward moving water masses for all scenarios, and b). the anomalous differences.

that decreases in are the dominant effect for controlling $H_o$.

Figure 6.11: Similar to Fig. 6.10 but for meridional mass transport $M_o$.

To directly analyse whether changes to $H_o$ are coupled more to variations in $M_o$ or , we plot the fractional differences of the latter two quantities to its time mean value, with equivalent changes in ocean heat transport. That is, $\delta M_o/M_o$ against $\delta H_o/H_o$, where for variable $X$, an overbar represents the time mean and $\delta X = X - \overline{X}$. These are shown in Fig. 6.12, with correlations with temperature difference $\Delta T$ (SH green, NH red) given by subplots a-d, and mass transport $M_o$ (SH black, NH blue) given by suplots e-h. From the scatterplots changes
in ocean heat transport are substantially correlated to changes in $\Delta T$, almost having a one to one relationship, with changes to mass transport playing a secondary role. The largest fractional changes to $\Delta T$ occur predominantly for the SH reaching about 60% at $4 \times \text{TRANS}$. In comparison, the NH has a smaller range in spread of about 20% across all scenarios.

For the scatterplots of $M_o$, the strong anti-correlations of a strengthening mass transport with decreasing ocean heat transport is visible, particularly for the SH (black stars). Correlations between $M_o$ and $H_o$ for the NH suggest smaller fractional changes overall to $\text{CO}_2$ concentrations below $3 \times \text{TRANS}$.

![Figure 6.12](image)

Figure 6.12: Scatterplots of changes between ocean heat transport and temperature difference $\Delta T$ (a-d), and meridional mass transport $M_o$ (e-h). NH points are represented by red and blue colours, whilst SH points are represented by green and black. The blue straight line represents the 1:1 gradient.

To study the origins of the decreases in $\Delta T$, we plot the anomalous latitudinal-depth profile for potential temperature across the Atlantic basin in Fig. 6.13, averaged between 300E-0N. The surface warming signal becomes stronger, spreading both polewards at depth with increasing $\text{CO}_2$, consistent with the SST warming shown in Fig. 6.1. Significant subsurface warming is observed at SH mid-latitudes, just northwards of the Southern Ocean boundary, and at high North Atlantic latitudes. For the latter region, the Polar Amplification warming effect of SST is convected into the ocean interior, warming up the ocean by approximately $0.5^\circ C$ at $2 \times \text{TRANS}$ up to depth of 1000m. At higher $\text{CO}_2$ concentrations this interior warming signal penetrates into spatial territories belonging to NADW (below 1000m) and AABW. No warming signals are observed for AAIW (100m-1000m) for the transient scenarios except for a temporary warming signal at $1.4 \times \text{TRANS}$. The lack of a clear signal may be due
to the relatively insensitivity of SST found near the Southern Ocean where AAIW originates. The increased northward Ekman transport in the region upwells cold interior water to the surface that eventually subducts and becomes AAIW.

Surprisingly there exists an interior cooling signal that is predominant in the $1.4 \times TRANS$ scenario, penetrating to depths between 2-5km. Similar signals are found in the later transient scenarios though these become weaker as warming signals from the surface start to penetrate into the interior. The depth at which these cold anomalies occur would suggest that any cooling affects originating near the surface could not possibly penetrate to such depths after almost a century of integration time. We conclude that this cooling signal is a result of CHIME’s ocean model not being able to conserve heat (Megann et al., 2010), with a globally integrated loss of $0.1^\circ C$ per century.

The gradual warming of the ocean interior is the main mechanism as to why the temperature difference between poleward and equatorward moving waters (and therefore ocean heat transport) decreases in a warming environment.

![Figure 6.13: Anomalous differences of potential temperature for the Atlantic basin region averaged between $300^\circ E-0^\circ E$.](image)

### 6.2.2 Dry Static Energy and Latent Heat Components

Atmospheric heat transport is decomposed into its dry static and latent heat transport components in Fig. 6.14, and its anomalous differences between each individual component in Fig. 6.15. As suggested by Bjerknes Compensation, atmospheric heat transport increases with $CO_2$, with different mechanisms depending on latitudinal region. At extra-tropical latitudes,
the increase in $H_a$ is primarily due to the increases in latent heat transport. In the tropical region, particularly for the SH, increases are due to the strengthening of DSE transport. At low CO$_2$ concentrations, namely $1.4 \times TRANS$ and $2 \times TRANS$, changes at extra-tropical latitudes are first identified as statistically significant. At $3 \times TRANS$ and $4 \times TRANS$ do signals in the tropics become significant, even becoming the dominant signal with magnitude changes of approximately 1.5PW near at 10°S.

The maximum atmospheric mass streamfunction for the SH Hadley Cell is shown in Fig. 6.16 for the entire transient CO$_2$ scenario integration spanning 130 years. Each individual transient climate scenario has been highlighted in a non-black colour and its time mean value given. Even with annual mean data, there is still considerable variation in the strength of the Hadley Cell by a one standard deviation value of $\pm 8.63 \times 10^9$ $kgs^{-1}$. There is however an underlying positive trend in the strength of $\Psi_a$, with an increase of 15% at $4 \times TRANS$ given a control mean value of $1.06 \times 10^{11}$ $kgs^{-1}$. As a comparison the strength of the NH Hadley Cell (Fig. 6.16, black dashed curve) shows an weakening trend over the same period. The opposing inter-hemisphere responses of each Hadley Cell are consistent with the different temperature gradient responses that act, in this model, to strengthen the circulation in SH and weaken it in NH.

A similar plot to Fig. 6.16 is established but for low level specific humidity, integrated between 700-1000 hPa, and between latitudes $0 – 30^\circ S$, shown in Fig. 6.17. Using the specific humidity values from $1.4 \times TRANS$ and $2.0 \times TRANS$ 10.0 and 10.7 g/kg respectively, the
Figure 6.15: Similar to Fig. 6.8 but for differences in meridional heat transport for the atmosphere (blue), latent heat (red), and dry static energy (blue) heat components.
anomalous global mean temperature difference between the two scenarios is $1^\circ C$, verifying that specific humidity increases at the Clausius Clapeyron rate of about 7% per Kelvin warming. The calculated control mean value of $q = 9.7 \text{ g/kg}$ results in an increase of 21% in value at $4 \times TRANS$.

By taking into account the increase in atmospheric mass transport of 15% and of low level specific humidity of 21% after roughly a quadrupling of $CO_2$ content, this would suggest a latent heat transport increase of $1.15 \times 1.21 = 1.39$, or approximately 40%. This magnitude is consistent with the rise of horizontal latent heat transport at $15^\circ S$ equalling roughly 1.8PW and 2.5PW for the control and $4 \times TRANS$ scenario respectively.

The increase in atmospheric heat transport for the SH tropics is the result of a larger increase in DSE rather than latent heat transport. The poleward moving heat term in expression of DSE ($c_p T + gz$) belongs to the gravitational potential term, since air parcels travel poleward aloft and equatorward below. One would suggest that an increase in DSE would be the result of an increase in the Hadley Cell height. Santer et al. (2003), Seidel and Randel (2006) and Zhou et al. (2001) observed an increase in the tropopause height by some tens of metres over the past few decades. By using the Hadley Cell streamfunction height (not shown) as a proxy for tropopause height, yielded no change in boundary height across all transient scenarios. This is due more to the CHIME atmosphere model not being able to resolve changes in pressure coordinates of a few tens of metres. We suggest that a more plausible explanation to the increase in DSE would be that the increase in circulation strength naturally advects a greater proportion of atmospheric mass at similar heights.

### 6.3 Inter-Basin Freshwater Transport

Anomalous SSS patterns shown in Fig. 6.3 reveal an increase surface salinity throughout the length of the Atlantic basin, intensifying with increasing $CO_2$ and suggesting increased anomalous net evaporation. The opposite is true for the entire North Pacific basin, with negative changes to SSS, corresponding to an increased anomalous net precipitation. The pattern however is not very clear for scenarios $1.4 \times TRANS$ and $2.0 \times TRANS$, only becoming clearly distinguishable at higher $CO_2$ concentrations. The changes between SSS for the Atlantic and Pacific basins are of the same magnitude, roughly 1psu change at $4 \times TRANS$. To satisfy this salinity imbalance suggests a net transport of freshwater from the Atlantic to the Pacific basin. A likely pathway for this freshwater transport is hypothesised to be through the Panama Isthmus of Central America, between $0 - 20^\circ N$; a full description of this mechanism was given in
Figure 6.16: The maximum strength of the Southern Hemisphere Hadley Cell as a function of time for the entire transient simulation (black solid). The colours mark the 10 year periods for specific transient periods of CO$_2$ concentration, with its mean value given by the arrows. The strength of the Northern Hemisphere Hadley Cell is given (black dashed) as a comparison.

Figure 6.17: Similar to Fig. 6.16 but for mean values of low level specific humidity averaged between 30°S-0°N and between 1000-700 hPa.

section 3.8.2.

To reiterate, from atmospheric dynamics, there are the Easterly Trade winds (travelling from the East) near the equatorial region, and the Westerlies (travelling from the West) at Mid-latitudes. As global mean surface temperatures increase throughout the transient scenario spin-up, so will there be an enhancement of the hydrological cycle with increased global evaporation and precipitation. This evaporation of water is picked up by the Easterlies within the
Atlantic and transported to the Pacific where it eventually enters into the ocean as precipitation. We hypothesise that this is a one way process, with freshwater not being able to re-enter from the Pacific to the Atlantic by the Westerlies due to Orographic features; the presence of the Rocky Mountains on the West Coast of North America and the Andes in South America acts as a natural barrier for the transport of moisture across the continents, precipitating any moisture travelling eastwards that eventually re-enters the Pacific via surface run-off.

To analyse this mechanism we look at the anomalous net spatial patterns of evaporation (E) minus precipitation (P) over the entire region. From previously Fig. 6.5 plots the spatial distribution of this quantity with the sign convention corresponding to positive (negative) values having a net increase in precipitation (evaporation). The spatial patterns, especially for $3.0 \times TRANS$ and $4 \times TRANS$ are in good agreement to the predictions of enhanced freshwater transport in the atmosphere (Lorenz and DeWeaver, 2007), whereby equatorial and high latitudes become wetter, whilst the Tropics become drier. There is a particularly strong signal of net freshening of surface water near the equatorial eastern pacific region. A corresponding net evaporation is seen at similar latitudes within the Central American basin, thus adding weight to the hypothesis that the net freshening seen within the former region originates primarily from the latter. An anomalous area of net evaporation develops south of the equator in both basins, though this is particularly pronounced for the eastern pacific. These regions describe sources of freshwater for the atmosphere which when coupled with strong sink regions directly northwards, suggests an overall northward shift in the ITCZ region within the transient spin-up scenario. There are both thus inferred regional and inter-basin mechanisms that are linked to the SSS anomalies.

### 6.3.1 Freshwater and Salinity Budget

We define our region of interest as the Eastern Equatorial Pacific (EEP), consistent with the terminology of Leduc et al. (2007). The boundaries for this area are described by the coordinates $200^\circ - 260^\circ$ longitude, $0^\circ - 20^\circ N$ latitude, and the first ocean layer depth of $5m$.

Fig. 6.18a illustrates the salinity evolution of the EEP describing a decreasing trend by about 1psu after $4 \times CO_2$. Fig. 6.18b shows the evolution of net E-P within the same region. The positive values for the control early transient scenarios describe a net evaporative region, consistent with the tropical climate. The decreasing trend in net E-P reveals that the region becomes progressively wetter throughout the transient run, completely reversing to negative values at $3.0 \times TRANS$ scenario. Fig. 6.18 (blue) illustrates the anomalous patterns of net E-
P, again highlighting the net increase in precipitation within the region. There is a significant increase in freshwater input between the 2nd and 3rd ‘TRANS’ scenarios compared with the 3rd and 4th, and is attributed to the northward shift of the precipitation rich ITCZ region within the EEP.

To find out whether the anomalous decreases in SSS seen in the EEP were controlled by the E-P anomalies we construct a salinity budget for the region to calculate how much freshwater input was needed for the model observed salinity changes.

We define the salinity equation to describe the EEP region can be expressed in the form

\[ M \frac{dS}{dt} = FS \]  

(6.1)

where \( M \) is the total mass of the surface layer, \( S \) is salinity, and \( F \) is net freshwater flux into the region (see Appendix, section 2.1.6 for a full derivation of the salinity equation).

Equation 6.1 is only valid for a closed system where seawater contained within the EEP region would be unmov ing, frozen constantly to the interactions of the atmosphere. In reality, the ocean is a highly evolving moving fluid, overturning in the vertical, never in constant contact with the atmosphere. We therefore rewrite the latter equation as

\[ M \Delta \frac{S}{\Delta t} = FS \]  

(6.2)

whereby \( \Delta S \) refers to the change in volume weighted mean salinity between a transient and control climatology, and \( \Delta t \) the overturning time it took for seawater to completely change within the EEP. This quantity was calculated from the divergence of seawater out of the region through horizontal mass transport and attained values of approximately 60 – 70 days. No significant correlation was found between the latter quantity and CO\(_2\) concentration. We note that this new approximation assumes that there is no net salinity transport into or out of the region from the flanks or from the bottom of the EEP. The freshwater input from the observed model salinity anomalies is illustrated in Fig. 6.19 (green) showing similar values to that computed from the anomalous net E-P (blue). Although there is a statistically significant net increase in moisture for all scenarios it is hard to quantify any differences since the standard deviation for each calculation overlap each other. Nevertheless, from comparisons of the two methodologies it is assumed that the changes in salinity within the EEP region are forced (to first order) by changes in net E-P at the ocean surface.

The evolution of mean values for salinity are illustrated in Fig 6.18a illustrating the decreasing trend with CO\(_2\) concentration.
6.4 Inter-basin Vs Regional Influences

Although there is seen an increasing input of precipitation through the evolution of $E - P$ across the EEP region, the source of this moisture transport is important as it underpins our central hypothesis that the SSS anomalies are forced by freshwater input originating from the Atlantic. If this is truly the case then much more freshwater would be incident from the eastern side of the EEP rather than from any other direction.

Net E-P is an atmospheric property, calculated from the transport of moisture at the region’s boundaries. We calculate net freshwater transport into the EEP by a third method. From our monthly mean data set we compute the column integrated moisture transport for the four sides of the region, namely North, South, West, and East. The imbalance of moisture entering and leaving served as an indirect measurement of net freshwater at the surface. The anomalous differences of freshwater transport calculated from atmospheric convergence is shown in Fig. 6.19 (red). From comparison of the latter calculation to the previous two already mentioned, the atmospheric convergence method reconstructs the actual net E-P model calculations with a fair degree of competency. The differences between the two methodologies can also be assumed to be the missing moisture transport effects by atmospheric eddies that are not temporally resolved with the monthly mean data. Nevertheless, it is well known that the Tropics are dominated by time mean activity such as the Hadley and Walker Circulations.
Figure 6.19: Anomalous changes of freshwater input into the Eastern Equatorial Region for three different calculations: net evaporation minus precipitation (blue), net moisture transport from atmospheric convergence (red), and indirectly from the SSS anomalies (green).

The freshwater fluxes for all sides and scenarios of the EEP region are plotted in Fig. 6.20. For the northern and southern sides, positive (negative) values constitute as northward (southward) moving freshwater transport, whilst for the eastern and western sides, positive (negative) values correspond to eastward (westward) transport. Thus, sources of freshwater into the EEP are from the east, north and south sides, whilst moisture leaves only from the western side. The convergence of moisture from the poleward sides represents the large equatorward transport of moisture within the northern hemisphere tropical Hadley Cell from the northern side; the northward transport at the south side represents the smaller contribution of freshwater convergence into the pacific ITCZ region. Transport from the flanks is both westward and consistent with the direction of the easterly Trade winds. For all scenarios, the westward side always transport more moisture away from the region than the contribution from the eastern side. The extra freshwater thus originates from the poleward sides and or via local evaporation within the region itself.

Figure 6.21 plots the anomalous differences of freshwater transport seen in the latter figure. From comparisons with all other sides, the change in the northern side represents the smallest changes, exhibiting no clear trends with $CO_2$ concentration. Though the EEP region is relatively small compared to the whole tropical region, the latter result is surprising as we would expect equatorward transport of latent heat (and therefore freshwater) from the north to increase with a warmer climate (Held and Soden, 2006). Since there is an overall anomalous
convergence of moisture into the region, we analyse the main sources of this freshwater, namely the southern and eastern sides. For $1.4 \times TRANS$ and $2.0 \times TRANS$, both sides exhibit similar increases in freshwater transport. At $3.0 \times TRANS$, the eastern side triples in magnitude its freshwater transport relative to the previous anomaly, when by $4 \times TRANS$ the eastern side transport stabilises to an increase of $1.8 \times 10^6 kgs^{-1}$. For the southern side, $3.0 \times TRANS$ represents no significant increase in freshwater transport compared to the previous anomaly climates. Only at $4 \times TRANS$ does freshwater transport more than double in size relative to $3.0 \times TRANS$. We relate the latter increase to a significant northward shift of the eastern pacific tropical ITCZ region which brings in much more moisture from the southern hemisphere.

Although the divergent transport of moisture from the western side is of similar magnitude to that of the previous two regions, there is no significant difference between the east-west sides of moisture transport, whereas there is in the north-south direction. This could be interpreted as moisture entering from the east will leave through the western side with no precipitation into the region. The net input would thus come predominantly from the south. Although this viewpoint has strong evidence we counter this argument stating that the freshwater leaving at the western side predominantly originates from the southern side, with moisture transport being directed westwards via the easterly Trade Winds.

The zonal distance of the EEP region spans $6.67 \times 10^6 m$ at the equator, reducing to
Figure 6.21: Anomalous differences of freshwater transport differences between transient and control climate scenarios represented in Fig. 6.20. Errorbars represent ±1 standard deviation. The x axis has similar labelling to that of Fig. 6.19.

$6.30 \times 10^6 m$ at $20^\circ N$. Since much of the moisture atmosphere lies between 700 - 100hPa (not shown) we compute the zonal mean westward velocity at the eastern side within this pressure range, estimated to be $\approx 5 ms^{-1}$. By taking the maximum parameterised value for the residence time of an air parcel within a cloud as 9 hours (Sanderson et al., 2010), this would suggest that the mean pathway of atmospheric moisture travelling from the eastern side would only traverse 2% of the zonal EEP region. From this simple calculation we can therefore assume that the majority of atmospheric moisture entering the EEP region from the eastern side will eventually precipitate into the region.

The comparisons of anomalous net freshwater transport from E-P calculations and from the eastern side by atmospheric advection are not of the same magnitude, with the latter values being slightly smaller. This implies that even if all the eastern input precipitates into the EEP region it would still not be enough to equal the surface net input across the whole area. The freshwater deficit by atmospheric convergence is thus cancelled by the increased freshwater input from the southern side, consistent with the northward propagation of the Pacific ITCZ region.

From the errorbars associated with Fig. 6.21, differences between control and transient scenarios only become significant for a doubling of $CO_2$ concentrations and above.
6.5 Unresolved issues

Although there is model evidence of an increase of atmospheric freshwater transport from the Atlantic to the Pacific through the Panama Isthmus with increasing atmospheric \( CO_2 \) concentration, there is insufficient evidence to prove that this increased source of freshwater is solely responsible for the observed model increases in SSS within the Pacific basin. From the analysis of the direction of freshwater transport into the EEP region the main inputs originate from the southern and eastern regions, although there is more of a larger influence from the latter region. The sources and directions of freshwater into the Pacific basin are a contribution of both inter-basin and regional mechanisms.

A particular puzzling feature of the observed model SSS anomalies is that the latitudinal extent of the increased anomalies within the Atlantic basin extends northwards towards the polar latitudes, implying an enhancement of the Atlantic Meridional Overturning Circulation (AMOC). This goes against IPCC predictions of a reduction in the AMOC strength with global warming due primarily to the enhanced stratification of the water column by the net increase of freshwater at high latitudes, suppressing the deep water formation regions that appear during northern hemisphere winter (Solomon et al., 2007). This issue is examined in Chapter 7.

Positive anomalous values of SSS have however been previously documented by Latif et al. (2000, Fig. 3b) and Thorpe et al. (2001). In their model analysis they found that anomalously high SSS’s developed in the Tropics were advected polewards, leading an increase in depth convection and a stabilisation of the overall weakening AMOC. The high poleward moving values of anomalous SSS’s in the North Atlantic is therefore not a new model observation.

As we shall see in the next Chapter, CHIME’s overall strength of the AMOC does decrease in the transient simulation.

Alternatively CHIME’s ocean isopycnal grid and mixing scheme has resulted in more heat and salt being trapped at the surface, with reduced mixing properties permeating these quantities into the interior, resulting in a sharper thermocline and pycnocline relative to climatological data (Fig. 2.9a and c). The lack of mixing between the near surface and ocean interior could effectively decouple, with a slow-down of the interior AMOC not being ‘felt’ by the upper ocean.
6.6 Changes in Heat Transport within EPcm Model

This section deals with results from the EPcm model to a doubling of $CO_2$. All plots associated with changes to individual variables can be seen in Appendix, section C. Within these parts extensive details of EPcm are recorded, pertaining to the model’s sensitivity to a range of factors such as holding $\Psi_a$ constant, holding $\Psi_o$ constant etc. $CO_2$ concentrations also go beyond the traditional extremes of a doubling of preindustrial values to give a larger perturbation (and therefore response) to the model. Although of interest, these features are not included in the main section since the analysis is seen as more of a model ‘curiosity’ than of something of new value. We therefore limit our investigation to the differences in heat transport after a doubling of $CO_2$, with an emphasis on the response of latent heat transport.

6.6.1 $2 \times CO_2$ results

Since a vast majority of studies focus on how the climate changes with a doubling of atmospheric $CO_2$ concentration, we take a similar approach and re-run the control integration experiment but with a prescribed doubled value of $CO_2$ at 560 ppm.

With a doubling of $CO_2$ in the model, global surface temperatures increase by 2.7K, which is in the range of most climate models (see IPCC AR4 report for more details).

The global moisture content scales as 6.4% per K warming; roughly in line with the predictions of Clausius-Clapeyron scaling of 7% per K warming.

Extra-tropical surface temperatures increase much more in the Extra-Tropics than in the Tropics, characteristic of the ‘Polar Amplification’ effect. This leads to a decrease in the temperature gradient that consequently drives decreasing trends in both ocean and atmospheric heat transports.

A surprising feature is that freshwater transport (or Latent heat transport in the atmosphere) reduces with increasing $CO_2$; in many present day climate models this quantity increases with $CO_2$ due to the increased moisture in the atmosphere from surface heating. We discuss later how we can ‘tweak’ the model to increase freshwater transport with increasing $CO_2$.

The salinity difference between tropical and extra-tropical surface waters slightly increases. Although in previous equations where we couple freshwater transport to the product of the ocean mass transport and the salinity gradient $F \approx \Psi_O \Delta S/S_O$, we would generally assume that an increase in salinity gradient would lead to an increase in freshwater transport. However
again due to the model’s parameterisation and tight coupling between the ocean and atmospheric mass transport, $\Psi_O$ decreases at a larger rate than the increase in salinity difference $\Delta S$, leading to the reduction in $F$ with increasing $CO_2$.

As the surface temperature warms and becomes more unstable the high latitudes starts to convect. This is behind the rate of increase in the hydrological cycle at 2.68% per K warming; much lower than Clausius-Clapeyreon scaling and in agreement with most climate models.

<table>
<thead>
<tr>
<th>2×CO₂ - Control experiment at equilibrium</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface temperature of Tropics</td>
</tr>
<tr>
<td>Surface temperature of Extra-Tropics</td>
</tr>
<tr>
<td>Global surface temperature</td>
</tr>
<tr>
<td>Atmospheric circulation strength</td>
</tr>
<tr>
<td>Global low level moisture</td>
</tr>
<tr>
<td>Atmospheric Moisture transport</td>
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<tr>
<td>Atmospheric heat transport</td>
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<tr>
<td>Oceanic heat transport</td>
</tr>
<tr>
<td>Total Heat transport</td>
</tr>
<tr>
<td>Hydrological cycle</td>
</tr>
</tbody>
</table>

Table 6.1: Changes to model variable results after a doubling of atmospheric $CO_2$ concentration.

### 6.6.2 The different responses of latent heat transport to an increase in $CO_2$

In section 3.5 we recall that we plotted the variations to certain quantities linked to heat transport as a function of varying the model’s parameters of atmospheric circulation strength $K_A$, and the water vapour feedback $\gamma$. Though the discussion back then concentrated on the control climate state, the figures in the section included data for the climate at 2×$CO_2$ concentration. It is these figures that we now concentrate on to analyse the different responses of latent heat transport to a doubling of $CO_2$. 


From figure 3.5c, we illustrate differences in response between the control climate state (blue line) and the double CO$_2$ state (red line) for the different multiple values of $K_A$. Out of all multiple values latent heat transport always decreases as CO$_2$ doubles. This is surprising since it has been shown in the previous sections, that latent heat transport should increase.

The calculation for latent heat transport is a simple product of atmospheric mass transport $\Psi_a$ and the specific humidity gradient, both shown in Fig. 3.5d and b respectively. The analysis of these two terms reveals that $\Psi_a$ decreases with CO$_2$ concentration, whilst the specific humidity gradient $\Delta q$ increases. The decrease in latent heat transport however results in the decrease of $\Psi_a$ being greater than the increase in $\Delta q$, negating this small increase in the formula of $L$.

To try and find a regime where the model increased with CO$_2$ concentration we repeated the analysis of varying $K_A$, but for the water vapour feedback parameter $\gamma$. The results are shown in Fig. 3.7. Latent heat transport was observed to increase with CO$_2$ for multiples of $\gamma$ less than unity, and decrease for multiple values greater and equal to unity. Through taking smaller increments to the multiple value of $\gamma$, we found that this turning point multiple was found to be at $\geq 0.7 \times \gamma$. The analysis of the variables $\Psi_a$ and $\delta q$, revealed similar behaviour experienced when varying $K_A$ i.e. decreasing atmospheric mass transport and increasing specific humidity gradient. This time however, the increase in the specific humidity gradient was greater than the decrease in the atmospheric mass transport with increasing CO$_2$.

This model attribute should not be confused with hysteresis whereby a model can exhibit more than one distinct equilibrium for the same starting conditions; here we are varying the parameter $\gamma$ each time thus differing our starting conditions for every integration period.

### 6.7 Chapter Summary

In this chapter we document a). the responsive features of meridional heat and freshwater transport in CHIME under an increase in atmospheric CO$_2$ concentration, and b). heat transport diagnostics in EPcm under a similar setting.

Large scale diagnostics in CHIME found that there was a steady increase in global mean near surface temperatures with increasing CO$_2$. Each hemisphere however displayed different responses, with the Northern Hemisphere exhibiting a reduction in its equator-to-pole temperature gradient reduction (Polar Amplification), whilst its southern counterpart experienced a temperature gradient enhancement. The latter feature has been linked to the upwelling of cold interior ocean water in the Southern Ocean that negates any near surface warming.
Changes to sea surface salinity showed an increase in salinification across the Atlantic basin, with the Indo-Pacific showing an increased freshening. Changes to net surface evaporation minus precipitation show an increase in net evaporation for the Tropics, and net precipitation for the extra-tropical region suggesting a poleward increase in freshwater transport. Near the equator, large spatially opposite dipole patterns of E-P infer a northward shift of the ITCZ particularly for the Pacific Ocean.

From the analysis of column integrated water vapour content it was found that CHIME followed the Clausius-Clapeyron rate. A similar analysis for precipitation revealed that CHIME increases by 1% per \( K^{-1} \), which is at the smaller range of increases calculated from the IPCC AR4 models (Held and Soden, 2006).

Changes to ocean, atmosphere and total heat transports behaved roughly to the views of Bjerknes Compensation, with changes to the ocean and atmosphere being roughly in opposite direction and equal magnitude such that their summation was roughly unchanged. Overall, the ocean heat transport tended to decrease whilst the atmospheric component increased. Any statistically significant trends however were only observed for \( CO_2 \) concentrations greater and equal to a tripling of control value \( CO_2 \), with the main areas of change being poleward of 40\(^\circ\) for both poles. This region reached down to 20\(^\circ\)S for the Southern Hemisphere but only for a quadrupling of \( CO_2 \). An analysis of ocean heat transport \( H_o \) as a product of meridional mass transport \( M_o \) and temperature gradient \( \Delta T \) surprisingly revealed that changes to \( H_o \) were primarily coupled to variations in \( \Delta T \), with \( M_o \) playing a secondary role. With decreases in ocean heat transport being coupled to decreases in temperature gradient, it was found that a large part of the ocean interior, similar in position to the location of the equatorward moving branch of the AMOC was warming. This supported evidence of the temperature of equatorward moving water masses increasing faster than the temperature of poleward moving water.

The internal decomposition of atmospheric heat transport into its latent and Dry Static Energy components revealed behaviour similar to Bjerknes Compensation, with changes to DSE and L being larger than that of changes to atmospheric heat transport, and of opposite direction. The increases in atmospheric heat transport are coupled to increases in latent heat transport at extra-tropical latitudes. Statistically this is only observed for a doubling of \( CO_2 \) or greater. At a quadrupling of \( CO_2 \), the largest changes to heat transport are observed within the tropical southern hemisphere region. An investigation of the Hadley Cell for the latter region showed a strengthening of the circulation by 15% after a quadrupling of \( CO_2 \). Similarly
low level values of specific humidity revealed an increase in values consistent with the Clausius-Clapeyron rate. The product of the two variables yielded increases of the same magnitude to explain the large changes in the Hadley Cell region. As a comparison, the strength of the Northern Hemisphere Hadley Cell exhibited a decreasing trend. The different hemisphere responses of the Hadley Circulation strength have been attributed to the different responses of the meridional temperature gradient across each region.

For the freshwater transport response in CHIME, the motivation originated from the anomalous SSS pattern of an increasingly salty Atlantic and freshening Indo-Pacific region, suggesting a net transport of freshwater from the Atlantic to the Pacific region via the atmosphere. This pathway was assumed to be through the Panama Isthmus, located between 0°−20° at Central America. A region was analysed in the Equatorial Eastern Pacific spanning 20° north-south by 60° east-west. Model calculations revealed that values of net evaporation minus precipitation was becoming smaller with $CO_2$, reversing to negative, implying the area becoming wetter. This was consistent with the magnitudes of decreasing SSS. By analysing the amount of freshwater through the 4 sides of the region it was found that the majority of freshwater into the region originated from the eastern and southern sides. The SSS decrease across the region thus resulted in an increase of net freshwater transport from the Atlantic into the Pacific through the Panama Isthmus (east side), but also from an inferred northward shift of the ITCZ region (south side). These results were only statistically significant after a doubling of $CO_2$ concentration.

Analysis of the EPcm under a doubling of $CO_2$ concentrate found that under its default parameters, latent heat transport decreased under $CO_2$ forcing, a counter response that has been seen CHIME and other IPCC class models. From analysis of the atmospheric circulation strength and the specific humidity gradient, it was found that the former decreased with $CO_2$, whilst the latter increased. The decrease in latent heat transport was apparent due to atmospheric mass transport decreasing at a faster rate than the increase in the specific humidity gradient. By ‘tuning’ the model, it was found that by decreasing the water vapour feedback parameter to a value below 0.7× the original value, latent heat transport would increase with $CO_2$. Within this regime, the increase in latent heat transport was responsible from the increase in the specific humidity gradient now being greater than the decrease in atmospheric mass transport.
CHAPTER 7

Changes to Ocean Mass Transport in an Increasing $CO_2$ Environment

7.1 Introduction

This chapter looks at how the ocean mass transport changes within temperature and salinity space, $M(T,S)$ (see Chapter 4 for theory) under increased atmospheric $CO_2$ forcing. Select latitudinal cross-sections are analysed, with specific relevance to changes in water mass properties.

NOTE – we remind the reader that the anomalous differences between a transient (increasing $CO_2$) and control scenario (fixed $CO_2$) is defined by “$CO_2$ value” $\times$ TRANS e.g. $3 \times CO_2$ would be equivalent to the differences between scenarios with a tripling of $CO_2$ relative to a control scenario at pre-industrial values. All transient scenarios are a 10 year average whilst the control scenario is 40 years.

7.2 Temperature and Salinity Anomalies

The Atlantic Meridional Overturning Circulation (AMOC) described in Chapter 4 and shown in Fig. 4.5D, describes the poleward transport of warm water near the surface, sinking at high latitudes due to the loss of buoyancy, and an equatorward return flow of cold water at depth.
We discussed in the previous chapter that the increases in Atlantic potential temperature between 1-2km below the surface were injected into the interior via the AMOC, whose positive temperature anomalies originated near the surface (see Fig. 6.13 from Chapter 6).

Figure 7.1 plots the anomalous cross-sectional distribution of salinity for the Atlantic basin. Salinity is taken as the zonal mean between longitudes $300^\circ - 360^\circ E$ The earliest permeating signal throughout all scenarios is the increase in salinity of the tropical near-surface waters (0-500m depth). Just as with the Atlantic basin temperature anomalies (Fig. 6.13), we propose that the positive salinity signal found at high northern latitudes are convected into the ocean interior by the AMOC, being observed only at $CO_2$ concentrations higher than triple the control value. At these same concentrations, a secondary signal emerges in the SH extratropical latitudes, that freshens water up to a depth of 1000m. This freshening signal extends into the equatorial region, where anomalous salty water sits above anomalous fresher water, resulting in a steeper pycnocline layer in the vicinity.

The next four figures show similar temperature and salinity cross-sectional anomalies for the Pacific and Indian ocean basins, averaged between the longitudes $150^\circ - 250^\circ E$ and $40^\circ - 110^\circ E$ respectively. We note that the vertical scaling has been changed from 0-5km to 0-2km depth resolution, due to much of the strongest anomalous signals being observed closest to the surface.

For the Pacific basin, the anomalous plots of temperature (Fig. 7.2) reveals a warming of
the near ocean surface (0-100m), with a larger signals originating within the NH than the SH, consistent with Polar Amplification (see Fig. 6.1). Around the Southern Ocean region, there is a negative temperature anomaly signal (cooling) that grows both in strength and spatial area with increasing $CO_2$. The pathway of the cooling signal is consistent with the AAIW region, leading to the assumption that AAIW cools with increasing $CO_2$. This may be explained as follows: considering that the equator to pole temperature in the SH increases, this results in stronger Westerlies over the Southern Ocean. In turn, this increases the equatorward Ekman Transport that sucks cool interior water to the surface, which in turn becomes the new colder AAIW; this is proved later in section 7.4.2. A sub-surface cooling signal is evident in the NH region, however the origins of this is not clear. The anomalous warm waters near the surface and cooler sub-surface waters below, for both hemispheres, act to steepen the thermocline layer and aid in the stratification of the water column.

The salinity signals of the Pacific basin are shown in Fig. 7.3. Apart from a localised near surface salinification signal between $20^\circ S - 0^\circ N$, the majority of the Pacific latitudinal extent displays a freshening signal throughout most of the first 500m of the ocean. The model observations are consistent with the results of Chapter 6 describing net evaporative Atlantic Ocean, and a net precipitative Pacific. The increased freshening signal of the AAIW is assumed to originate from the anomalous net precipitation found in the Southern Hemisphere. Since the latter region is the source region of AAIW, the freshening signal at the surface would therefore be inherited by AAIW as one of its properties.

Figure 7.2: Similar to fig. 7.1 but for temperature anomalies and for the Pacific Ocean region.
Figure 7.3: Similar to fig. 7.1 but for the Pacific Ocean region.

The temperature anomaly signals for the Indian Ocean basin (Fig. 7.4) are quite similar to the of the Pacific in that they both show a near surface warming anomaly followed by a net cooling signal below; the abrupt end to all signals near the equator is due to the presence of land. For changes to salinity, apart from a slight increase in salinity at $1.4 \times TRANS$, there is an overall freshening trend that is consistent with the AAIW pathway.

Figure 7.4: Similar to fig. 7.1 but for temperature anomalies and for the Indian Ocean region.

In the next section we analyse how the anomalous changes of temperature and salinity for each of the major basins impact on the presentation of data given by $M(T,S)$ analysis.
7.3 T-S Anomalies

The changes to the temperature and salinity profiles of the ocean with $CO_2$ concentration will undoubtedly change the distribution of $M(T,S)$ plots of individual latitudes. This section aims to analyse these changes. Due to the amount of data already present, we use a 2 dimensional T-S diagram instead of the usual $M(T,S)$ diagram for two reasons: firstly, to simplify the data by concentrating solely on the distribution and not the magnitude, and secondly, for better visualisation. As a further simplification, only two scenarios are used, namely the control and $4 \times TRANS$ (quadrupling of control value $CO_2$). This extreme choice of scenario was chosen to allow for the largest shifts of water masses in T-S space to be observed, allowing for minimal overlap of water masses which were often observed for lower $CO_2$ transient scenarios. Figure 7.6 shows individual T-S diagrams identical to the latitudes used in Fig. 4.13 from the previous chapter. Blue colours represent the T-S diagram for $1.4 \times TRANS$ whilst red colours represent $4 \times TRANS$. The global T-S map is shown on the left column whilst the Atlantic portion is shown on the right. A complete description of $M(T,S)$ plots and its basic interpretations are given in Chapter 4.

For the latitude $40^\circ N$ (Fig. 7.6A and B) there is a clear separation of water masses belonging to the Atlantic (right, salty) and Indo-Pacific (left, fresher) regions, each with unique characteristic changes. The Indo-Pacific sees a shift towards the top left of the diagram, particularly for its top branch, indicating warming and freshening of near surface waters. In contrast, the Atlantic sees a shift towards the top right of the diagram that reveals a warming
and salinification of waters throughout the entire water column.

At 24°N (Fig. 7.6C and D) many of these changes still hold. The anomalous cooling and freshening signals found in the Indo-Pacific basins results in water mass signals being shifted downwards along isopycnal surfaces.

For the latitudes 0°N and 20°S (Fig. 7.6E, F, G and H), the characteristic ‘trunk’ of the AAIW signal (thin localisation of water masses approximately between 5 – 15°C) shifts downwards along isopycnal surfaces consistent with the freshening and cooling signals observed in the above cross-sectional figures of temperature and salinity for each basin. Likewise for NADW, there is a shift upwards along isopycnal lines, resulting from the warming and salinification of the interior Atlantic basin (see Fig. 7.6F for details).

At 34°S (Fig. 7.6I and J) the cooling and freshening signal of the AAIW persists. Near surface waters for both the global and Atlantic regions follow a general warming of water masses, with a slight freshening shift.
7.4 Mass Transport

7.4.1 Changes to the Atlantic Meridional Overturning Circulation

As discussed above, the AMOC describes a single overturning circulation cell that stretches the full latitudinal extent of the Atlantic basin, advecting water northwards at near surface depths with a return flow in the ocean interior. The sinking of dense waters in CHIME occurs at high Northern latitudes between 50 – 60° N in CHIME, consistent with the Labrador and GIN (Greenland, Iceland, and Norway) Sea region where deep convection is known to occur. The strength of the AMOC is controlled primarily by the ‘push’ of downwards water from deep convection in the high latitude NH, and the ‘pull’ of deep water to the surface from Ekman suction in the SH. This next section aims to analyse the changes to the AMOC circulation and offer theories to explain the model outputs.

We calculate the meridional volume streamfunction of the Atlantic ocean using Eq. 4.2 used in Chapter 4, and highlighted in Fig. 4.5D. The maximum strength of the AMOC is taken as the maximum value of the volume streamfunction which we denote \( \Psi_{\text{max}} \). Fig. 7.7 plots \( \Psi_{\text{max}} \) for the transient run scenario which increases \( CO_2 \) concentration by 1% per year at year 60 up to year 200.

The data uses a 12 month smoothing filter to remove the seasonal cycle. The first 40 years shows a temporary upward trend in AMOC strength to approximately 21 Sv at year 100. After this period there is gradual weakening of the AMOC for the next 100 years. An approximated linear trend for this period (black dashed line) is a decrease of 5 Sv per century, or a 24% decrease over the same period. This model result is within the middle range of the IPCC AR4 predictions of an AMOC decrease in strength by 0-50% by the end of the century (see section 1.5.5 for more details.). To analyse why the strength of the AMOC decreases in the last century we examine the net surface heat flux at high northern latitudes. Deep convection occurs within the Labrador Sea and off the coast of Greenland when there is a large net surface heat flux out of the ocean and into the atmosphere, leading to the formation of very dense waters and the enablement of vertical motion. This process is most active during the winter months when the temperature difference between the warm ocean and cold atmosphere is at its greatest. The amount of heat lost by the ocean at these high latitudes can be used as a measure of the strength of the AMOC. An area was chosen to contain the deep convection regions of the Labrador Sea and territories south of Greenland and north-west of Scotland; this we term the Deep Convection Region (DPR) and has coordinates 296.25 – 352.50° E by
55.00 – 67.50°N. The area weighted net heat flux at the ocean’s surface for the DPR region was computed for every February month for the transient \(CO_2\) period, the results of which is plotted in Fig. 7.8 (blue curve). A 10 year smoothing filter is used on the data to remove much of the variability. As a comparison the strength of the AMOC is shown (red curve) with its magnitude multiplied by -9.5 to be rescaled with the blue curve. Between years 60-100 net surface heat flux loss increased from almost 180 Wm\(^{-2}\) to 200 Wm\(^{-2}\). This would have aided in the production of dense water and a corresponding strengthening of the AMOC that was observed over the same period. After year 100 we see a gradual weakening of net surface heat loss to approximately 150 Wm\(^{-2}\) at year 173. This decrease comes about due to the warming trend of the atmosphere associated with increases in atmospheric \(CO_2\) levels; since the net surface heat flux loss is proportional to the temperature difference between the warm ocean and colder atmosphere, the gradual warming of the latter reduces the temperature difference between the two media, resulting in the reduction of net heat loss and dense water production that matches well with the corresponding reduction in strength of the AMOC. A cross correlation between net surface heat flux and the strength of the AMOC (not shown) revealed that the highest index between the two time series’ was when the AMOC strength lagged behind net surface heat flux by 5 years.

A puzzling result of the decrease in AMOC strength is that CHIME shows a northward propagation of positive SSS anomalies up to 70°N (shown in Fig. 6.3), that can be seen as consistent with a strengthening of the AMOC (Manabe and Stouffer, 1988). We shall however argue that the high SSS anomalies in the Atlantic can still be congruent with the model observations of a weakening AMOC.

To examine this apparent paradox we employ the use of Hovmöller plots. Figure 7.9 plots the zonal mean values of anomalous SSS, SST, E-P and sea surface density as a function of longitude (x axis) and time (y axis) across the Atlantic basin only. We note that the full longitudinal extent of the Atlantic basin was used to compute zonal mean values, compared to the average between 0 – 60°W that was employed in Fig. 7.1. From the examination of Fig. 7.9A high SSS anomalies develop in the sub-tropics at about 10°N and propagate northwards up to 50°N at a velocity of \(\approx 1 \text{ cms}^{-1}\). This value was calculated from the slope of the thick black line (speed = distance/time) and found to be consistent with the explicit calculations of zonal mean meridional surface velocity (not shown). The origins of these high SSS anomaly signals agree well with the persistent strong net evaporation signal observed between 10 – 20°N, illustrated in Fig. 7.9C, and its corresponding weaker net
evaporation signal up to 50° N. Between 50 – 60° N a negative SSS signal is observed, in line with net precipitation over the same region. Though we mentioned earlier that the anomalous SSS signals for the North Atlantic reached up to 70° N, this negative signal is weighted towards the region of the Labrador Sea where its large but localised negative SSS signal is enough to counter any weaker positive signals when involved in a zonal average. The onset of ‘strong’ positive SSS anomalies can be seen to start at around year 120, 20 years later than when the AMOC first starts to weaken in strength, suggesting that the formation of these high positive SSS anomalies are a consequence of the weakening of the AMOC. An explanation for this mechanism can be as follows: a weakening AMOC at year 100 would reduce the meridional velocity near the surface. Northward moving surface water parcels travelling across the net evaporative tropics would therefore be exposed to the atmosphere for a longer period of time, resulting in higher SSS anomalies. These anomalies would be re-enforced by the net evaporative region between 10 – 50° N.

The Hovmöller plot of surface density anomalies was computed using the equation of state for the ocean with inputs from surface temperature, salinity and pressure as variables. It was documented by Thorpe et al. (2001) and Latif et al. (2000) that the northward advection of high SSS anomalies developed in the Tropics aided in deep convection and a stabilisation in the overall decrease of the AMOC. Although in the CHIME we document a similar advection of high SSS anomalies, these do not penetrate past 50° N (according to the Hovmöller plots) to affect the deep convection sites. Additionally, an examination of surface density anomalies reveals that after year 100, surface density is overall lower across much of the Atlantic basin. This would suggest that since density is a function of salinity and temperature (taking pressure as constant at the surface), the effects of increasing SST more than compensate for any changes in SSS made by net evaporation/precipitation. From this evidence we state that no stabilisation of the AMOC in CHIME will originate from the transport of high SSS anomalies into the deep convection region.

7.4.2 Changes to mass transport in density space

The changes in temperature and salinity to the global ocean basins have an effect on the density range of seawater. From the use of T-S analysis, these have been identified as along-isopycnal evolution of water masses. The densities of water masses should change, and as shown from the weakening of the AMOC, the transport strength of certain water masses should also change; this section aims to analyse these changes.
Figure 7.10 plots the volume transport in Sv as a function of $\sigma_t$ for the specified latitudinal cross-sections described above (rows). To reduce the amount of curves shown, only data from the control run (blue curve) and at $4 \times CO_2$ (red) are illustrated. The resolution used in density space was $\Delta \sigma_t = 0.3$. Plots for the global integration of volume transport are shown in the first column, with components belonging to the Atlantic and Indo-Pacific region given in the second and third columns respectively; the corresponding M(T,S) is illustrated in fourth column but only for the control run scenario. Positive (negative) magnitudes relate to northward (southward) flow.

As explained in Chapter 1, density can be used as an approximation to vertical height in the ocean, with small (large) values of $\sigma_t$ corresponding to shallow (deep) depths. Water masses near the surface influenced by wind driven processes tend to have a much more diffuse distribution on scattered across isopycnal surfaces. Slower, interior flows have less mechanical mixing and therefore tend to have a sharper distribution of water masses on isopycnal surfaces.

The clearest water mass signature is that of NADW, corresponding to a negative peak of $\sigma_t \approx 37 \pm 0.5kgm^{-3}$ found at all given latitudes. No signal is found in the Indo-Pacific since NADW is exclusively found in the Atlantic basin only. For latitude $40^\circ N$ (row A), the balance of seawater transport in the Atlantic is between northward flowing near surface waters, and the southward flowing NADW.

At $24^\circ N$ within the Indo-Pacific region (Fig. 7.10Biii) the majority of water masses that balance light northward flowing waters and dense southward flowing waters are both lighter than the $\sigma_t = 34 kgs^{-1}$ surface, which when compared to its water mass distribution on an M(T,S) plot, describes a wind driven gyre circulation. Although the Atlantic branch of water masses on the M(T,S) map show the similar gyre circulation signal to that of the Indo-Pacific, it is surprising that this is not reflected in the corresponding vertical density distribution (Fig. 7.10Bii) given that the latter region should in theory have a stronger wind driven circulation due to its narrow basin width.

Flowing directly below NADW in physical space, the signature for AABW is a northward flowing water mass situated at a slightly heavier density of $\sigma_t \approx 37.2 \pm 0.5kgm^{-3}$. Originating off the coast of Antarctica, its maximum latitudinal extent (in CHIME) does not penetrate more than $24^\circ N$ and so is not observed at $40^\circ N$. AABW should be observed flowing in both regions however no signal is observed in the Atlantic. To explain this discrepancy we assume that the flow of AABW for each basin is proportional to its longitudinal extent; the flow in the Atlantic would be about a fifth the amount flowing in the Indo-Pacific. Due to the close
proximity between AABW and NADW in density space, the small flow signature of the former would therefore be consumed by the latter, leading to no distinct AABW signature for the Atlantic basin.

AAIW is formed near the surface of the Southern Ocean region where it flows northward at a shallow depth of between 500-1000m to just over 0°N (only plots C, D, and E). On the M(T,S) map its signature is the vertical ‘trunk’ of water masses between \( \sigma_t = 33 - 36.5 \text{ kgm}^{-3} \) giving it a larger density range than for either AABW or NADW. Its mass transport is largest for the Indo-Pacific region due to its larger cross-sectional area compared to the Atlantic, and decreases with latitude.

As a comparison for the major water masses described above within a control climate environment to one after a quadrupling of \( CO_2 \), we observe an overall reduction in the volume flux of NADW and AABW, and an increase in the flow of AAIW. Our analysis for this is given below:
We compute the amount of volume flux for each water mass as an integral of mass transport within a certain density range, i.e., \( M_{wm} = (\sigma_t) d\sigma_t \), where subscripts \( wm \) signify water mass. The density range for each of the major water masses were empirically derived from the plots of Fig. 7.10. Although the latter figure only shows results for the control and \( 42 \) environment, we also calculate \( M_{wm} \) for each of the other transient scenarios. Figure 7.11 plots the volume transport for the water masses of NADW, AAIW and AABW respectively, as a function of \( CO_2 \) concentration relative to the control climate reference of 280 ppm. Each different coloured curve represents the different latitudinal cross sections taken, with the convention that positive (negative) volume transports represents northward (southward) flow.

For NADW whose formation region is at high northern latitudes in the Atlantic, the zonal cross section of \( 40^\circ N \) is nearest to this region and best mirrors the original flow rate of NADW. At this latitude, almost all of the southward moving water of the AMOC is made up of NADW. The decrease in transport of NADW is therefore well correlated (in magnitude) to the reduction in overall AMOC strength shown in Fig. 7.7. The changes in magnitude of NADW for zonal cross-sections further southward infer that there is a net production of this water through to \( 0^\circ N \), with a net destruction southwards up to \( 34^\circ S \).

For AAIW, its closest cross section to its formation region of the Southern Ocean is \( 34^\circ S \). At this latitude, flow rate increases from 43 Sv to 59 Sv over a 140 year period. We suggested before that the increase in AAIW was due to an increase in westerly windstress over the Southern Hemisphere that resulted in an enhanced upwelling of cold water and northward Ekman transport that fed AAIW. A calculation of the latter however revealed Ekman transport increased by only 6 Sv over the same 140 year period. Due to the difference between the \( 34^\circ S \) M(T,S) section and the \( 55^\circ S \) zonal windstress calculation, we cannot conclude with confidence whether or not the increase in AAIW is due primarily to the increase in zonal wind stress since production of AAIW through mixing could have taken place between the two latitudes, a process which is true between the latitudes \( 34 - 20^\circ S \) by almost 30 Sv. Between \( 20 - 0^\circ S \) there is a dramatic destruction of AAIW with flow rates dropping to values even below that at \( 34^\circ S \).

The production and flow of AABW is comparatively well behaved; having been formed off the coast of Antarctica volume transport decreases with latitude away from this region, with the majority of latitudinal cross-sections reducing AABW transport with increased \( CO_2 \) (except \( 24^\circ N \) and \( 40^\circ N \)). The reduced volume transport of AABW as a function of latitude for all \( CO_2 \) concentrations means that over its northward propagation, there is an eventual
destruction of its water mass through mixing. Since AABW is formed through the formation of very dense surface water through brine rejection from ice formation, a reduction of AABW would suggest an overall reduction in Antarctic ice formation during the Southern Hemisphere winter years. An examination of the annual mean area weighted ice cover fraction for all territories southward of −57°S (not shown) revealed that for the entire 140 year transient CO$_2$ scenario, ice fraction decreased from a maximum of 40% to 38% over the last 100 years. It is worth noting that the latter variable was well correlated to the strength of the AMOC, with a cross-correlation between the two variables being strongest when the AMOC lagged behind ice-cover fraction by two years. A calculation of the August net surface heat flux for the same region and transient CO$_2$ scenario (not shown) produced a linear trend from about −100 W m$^{-2}$ to −90 W m$^{-2}$ for the entire 140 year period, inferring a reduced loss of heat from the ocean to the atmosphere. Both of these reductions are consistent with the decrease in AABW transport shown in Fig. 7.11C.

7.5 Summary

We study several properties of how the ocean changes with increasing CO$_2$ concentration, using a mixture of analysis techniques. For all the major basins near surface temperatures increase for the majority of latitudes, with the strongest warming found at high NH latitudes, whilst high SH latitudes observe a cooling signal. The Atlantic basin interior begins to warm slowly as the southward branch of the AMOC brings in the anomalously warm waters from the surface. For all major basins AAIW cools, a consequence most probably resulting from the increased upwelling of cold interior water within the Southern Ocean from enhanced northward Ekman Transport.

Salinity anomalies reveal an increase in salinification for the Atlantic basin northwards of the Southern Ocean regime, with considerable increases found within the southward branch of the AMOC. The Pacific and Indian Oceans share a freshening signal across all latitudes, providing inferring evidence of a net transport of freshwater from the Atlantic into the Indo-Pacific region.

T-S plots reveal that the Atlantic becomes warmer and saltier, whilst the Indo-Pacific becomes warmer and fresher. This trend is much more apparent for the NH latitudes than within the SH. AAIW water masses tend to get fresher and colder.

The strength of the AMOC was examined for the complete transient CO$_2$ scenario increasing at 1% per year for 140 years. It was found that for the last century, the AMOC
decreases from a maximum of about 21 Sv to 16 Sv; a prediction well within numbers given by the IPCC AR4. To study the decrease in AMOC strength a region of analysis was chosen that included the deep convection sites of the Labrador Sea and territories south of Greenland and westward Scotland. During NH winter, the difference in temperature between the warm ocean and colder atmosphere results in a loss of net surface heat flux from the ocean to the atmosphere. This in turn cools the ocean surface and increases density of the top layer, destabilising vertical stratification and leading to downward convection. It was found that the loss of net surface heat flux during February decreased from $-200 \text{ Wm}^{-2}$ to $-170 \text{ Wm}^{-2}$ for the last 100 years, consistent with a warming atmosphere relative to that of the ocean from anthropogenic forcing. Indeed a cross-correlation between the two variables found correlations was strongest when net surface heat flux led by 5 years, verifying our description that AMOC strength is controlled by net surface heat flux loss during winter.

A consequence of the slowing down of the AMOC was the formation and northward advection of positive SSS anomalies for the entire Atlantic basin. From the use of Hovmöller plots these positive SSS anomalies were found to originate between the ‘high’ net evaporation band between 10 – 20°N. Using the slopes on the Hovmöller plot these signals were found to propagate northwards by approximately 2\text{cms}^{-1}, consistent with the explicitly calculated meridional mean sea surface velocity. As the observed start of the AMOC weakening is at year 100, the propagation of high SSS anomalies is only seen at year 115, establishing that the weakening of the AMOC precedes any changes to SSS. We theorise that as the AMOC slows down, the northward propagation of high SSS water developed within the Tropics has much more exposure time to the atmosphere, sustaining these signals as they propagate northwards. Thorpe et al. (2001) and Latif et al. (2000) argued that the propagation of high SSS signals would act as a negative feedback to the slowing of the AMOC, aiding to restart deep convection. However with CHIME, although there is a propagation of high SSS anomalies, the increase in surface temperature from the warming climate overrides these changes in salinity such that surface density in fact becomes lighter, weakening any effects of deep convection.

Plots of volume transport as a function of density were examined for five latitudes: 34°S, 20°S, 0°N, 24°N and 40°N. Depending on latitude, distinct peaks in volume transport occurred at specific density ranges, corresponding to the major water masses of the ocean, namely NADW, AAIW and AABW. It was found that the volume flux of NADW decreased from about -22 Sv to -16 Sv for the entire 140 year transient period (calculated at 40°N, mirroring a similar reduction in AMOC strength. The transport of AAIW saw an increasing
trend from about 43 Sv to 59 Sv over the same period, calculated at 34°S. We attributed this trend to the increase in northward Ekman transport at 55°S, however the latter only accounted for only an increase of 6 Sv, with the difference being made up by mixing of water into the density range of AAIW. Computed at 34°S, AABW saw a decrease from about 20 Sv to 8 Sv. Since the formation of AABW is linked to the accumulation of dense surface waters from brine rejection due to ice formation, and the loss of net surface heat flux, we studied these two mechanisms for the entire region southward of 57°S. It was found that area weighted ice cover fraction decreased from a maximum of 40% to 38% over the last 100 years. Surprisingly, ice cover fraction was well correlated to the strength of the AMOC even though the coast of Antarctica is separated from the Atlantic by the Southern Ocean. Net surface heat flux saw a reduction from about $-100 W m^{-2}$ to $-90 W m^{-2}$ for the entire 140 year period, inferring a reduced loss of heat from the ocean to the atmosphere. Both reductions to these two variables would be partial to explaining the weakening of AABW production and transport.

In all, changes at the ocean-atmosphere interface are not only able to change the ocean’s internal composition of water mass, but also alter the circulation of its fundamental water masses.
Figure 7.6: T-S diagrams for specified latitudes. The control run scenario is shown in blue whilst the $4 \times CO_2$ scenario is represented by red. The global cross-sectional integral is given in the left hand column whilst signals belonging solely to the Atlantic basin are given in the right hand column.
Figure 7.7: The maximum strength of the AMOC $\Psi_{\text{max}}$ for the transient scenario. $CO_2$ increases at 1% per year starting at year 60 (280 ppm) and continues up to year 200 (1120 ppm). The black dashed line represents a decreasing linear trend of approximately a 5 Sv decrease over the next century starting from year 60.

Figure 7.8: Area weighted net surface heat flux over the Deep Convection Region for February months (blue) within the transient $CO_2$ scenario forced at a1% increase per year. A 10 year running mean is used for the latter time series to remove some variability. The maximum strength of the AMOC is added (red line), rescaled by multiplying by -9.5.
Figure 7.9: Hovmöller plots for the Atlantic basin of A). sea surface temperature anomalies, B). sea surface salinity anomalies, C). net evaporation minus precipitation anomalies, and D). sea surface density anomalies. Differences are between a 40 year mean control climate state and a transient $CO_2$ state that increases by 1% per year for the next 140 years. Annual mean data is used.
Figure 7.10: Volume transport as a function of $\sigma_t$ for the latitudes A. $40^\circ N$, B. $24^\circ N$, C. $0^\circ N$, D. $20^\circ S$, and E. $34^\circ S$. Calculations for the global cross-sectional integral are given in column i., whilst the Atlantic and Indo-Pacific regions are shown for columns ii. and iii. respectively. Blue (red) lines indicated the control ($4 \times CO_2$) climate respectively. The corresponding M(T,S) plot is shown in column iv. though only for the control climate; black curved lines represent isopycnal surfaces whilst the colour convention for water masses
Figure 7.11: Volume transport for A). NADW, B). AAIW, and C). AABW as a function of $CO_2$ concentration (given as the multiple of the control climate value of 280 ppm). Coloured lines indicate different latitudinal cross sections. Differences in volume flux between different latitudes indicate either a net formation (destruction) of that particular water mass. Units are in Sv. Control climate variables are from a 40 year mean, whilst transient $CO_2$ scenarios represent a 10 year mean period centred a particular $CO_2$ concentration.
Summary

Throughout this thesis we have investigated the roles of meridional heat and fresh water transport, concentrating on three primary areas:

1. sensitivity of heat and freshwater transport to model numerics (Chapters 3 and 4)
2. coupling between heat and freshwater transport (Chapter 5)
3. changes to heat and freshwater transport to increasing CO$_2$ concentration (Chapters 6 and 7)

For the analysis of the above points the use of three climate models were employed; two being fully coupled climate models (HadCM3 and CHIME [Chapter 2]) of IPCC class resolution, and a simple two-box model EPcm (Chapter 2).

8.1 Comparison of Models Simulations in a Control Climate Scenario

In Chapter 3 we examined the sensitivity of meridional HT to the choice of ocean vertical coordinate system. The models HadCM3 and CHIME share similar qualities, employing identical atmospheric models (HadAM3) and identical ocean horizontal resolution southwards of 55°N. The primary difference is that whilst HadCM3 uses constant depth for its ocean vertical coordinate system, CHIME employs an isopycnal coordinate system in the vertical. Coupled with
the specific mixing schemes associated with each vertical coordinate system, comparing both models offers the perfect test bed to observe how heat and freshwater transport are dependent on the choice of ocean model architecture.

A central idea that is used extensively originates from the work by Stone (1978) that states that to first order, the distribution of total meridional HT can be seen as relatively insensitive to any changing processes that occur within the ocean or atmosphere. We hypothesised that since CHIME and HadCM3 had identical atmospheric models and forced with the same initial conditions and pre-industrial CO$_2$ levels, inter-model differences of total heat transport would be negligible, and that any large amplitude changes to either ocean or atmospheric HT would be of opposite but equal magnitude so as to conserve total energy; the latter argument is known as Bjerknes Compensation (Bjerknes, 1964) and is another central idea to the comparison between HadCM3 and CHIME.

![Figure 8.1: Differences in the meridional HT (CHIME - HadCM3) for components belonging to A). total, ocean and atmosphere, and B). atmosphere, dry static energy and latent heat components. Adapted from Fig. 3.10.](image)

In the control simulation, the comparisons of total heat transport yielded differences of approximately 0.2PW with CHIME being slightly larger. To first order, total HT for both models can be seen as identical to each other (Fig. 8.1A). Differences to the ocean and atmospheric components were of opposite but similar magnitude of 0.5 PW, with CHIME having a slightly weaker ocean and stronger atmospheric contribution respectively, consistent the ideas of Bjerknes Compensation. A similar procedure was carried out to the atmosphere’s HT components of latent and dry static energy transport. Inter-model differences between the
latter two revealed larger differences than what was observed for even the ocean or atmosphere media (Fig. 8.1B). The main results of the analysis are summarised below:

- Atmospheric HT was decomposed into its Dry Static and latent heat contributions for both models
- An examination of the model differences revealed changes spanning more than 1PW at 7.5°N, larger than changes due to the ocean or atmosphere alone
- at this latitude it was found CHIME’s equatorward latent heat transport and poleward dry static energy transport were both reduced in comparison to HadCM3
- by stating that latent heat transport carried was proportional to the strength of the Hadley Cell circulation and the specific humidity difference between poleward and equatorward moving air, it was found that the former quantity decreased by 52% whilst the latter increased by 9%
- differences in SSTs between the two models revealed that CHIME had warmer mid-latitude regions within the NH concentrated near the major western boundary current regions
- this feature acted to reduce the equator-to-mid-latitude temperature gradient that resulted in the decreasing Hadley Circulation strength

As SST couples the ocean to the atmosphere, we have observed how slight model changes can have very large changes to HT within the climate system, even at fixed CO₂ concentrations.

To compare the freshwater transports between the two models within a control climate (other than the meridional latent heat transport already discussed) net fluxes of evaporation minus precipitation over all the major ocean basins were calculated. It was found that for both models the Atlantic and Indo-Pacific regions were net evaporative, whilst the Southern Ocean and land cover were net precipitative. Model differences however identified that the Atlantic basin in CHIME as being slighter more net evaporative than HadCM3, whilst its Indo-Pacific region was slightly less evaporative; both changes were on the order of 0.05 Sv (see Table 8.1).

This implied an anomalous net transport of atmospheric freshwater from the Atlantic into the Indo-Pacific region. It was hypothesised that the likely pathway was westwards through the Panama Isthmus of Central America (Broecker, 1991). Direct calculations of zonal freshwater transport through a latitudinal cross-section between 0°N - 20°N at the Central American
Table 8.1: Net surface freshwater flux for the Atlantic (at), Indo-Pacific (ip), Southern Ocean (s), and land (l) contributions. Northward of 70°N $F_s$ constitutes both land and sea areas. Lateral fluxes from continental runoff have been excluded. Data taken from years 80 - 119 of a control climate scenario. Both the Atlantic and Indo-Pacific regions are defined between latitudes 30°S to 70°N, whilst the Southern Ocean is defined by the area southward of 30°S. Units are in Sv. Red (blue) colors indicate key net evaporation (precipitation) values. Adapted from Table 3.2

<table>
<thead>
<tr>
<th>Model</th>
<th>$F_{at}$</th>
<th>$F_{ip}$</th>
<th>$F_s$</th>
<th>$F_l$</th>
</tr>
</thead>
<tbody>
<tr>
<td>CHIME</td>
<td>1.15</td>
<td>0.86</td>
<td>-0.76</td>
<td>-1.25</td>
</tr>
<tr>
<td>HadCM3</td>
<td>1.11</td>
<td>0.91</td>
<td>-0.75</td>
<td>-1.28</td>
</tr>
<tr>
<td>CHI-Had</td>
<td><strong>0.04</strong></td>
<td><strong>-0.05</strong></td>
<td>-0.01</td>
<td>0.03</td>
</tr>
</tbody>
</table>

cost yielded values of $-0.43 \pm 0.03$Sv and $-0.39 \pm 0.07$Sv for HadCM3 and CHIME respectively. The calculations however show HadCM3 has a larger westward freshwater transport by 0.04 Sv relative to CHIME, the opposite to what was expected. Although the size of the error bars does open the possibility that CHIME could have a stronger freshwater transport, this may be ‘stretching the truth’ using statistics, especially when the standard deviations are approximately one order of magnitude larger than its mean variable. Different latitudinal ranges were also tried (20°S-20°N, 10°S-20°N, 0°N-10°N), though all resulted with similar conclusions. Due to the small model differences needed to be calculated, both models’ internal variability proved too large to draw any meaningful conclusion from these results. Given the results, our original hypothesis was proved invalid. We however feel that the Atlantic-Pacific freshwater transport mechanism is not ‘wrong’, but merely needed a larger model difference to differentiate time mean calculations from natural variability.

In chapter 4 we further investigated differences between CHIME and HadCM3 within a control climate. Calculations relating to ocean heat transport led to the mapping of ocean meridional mass transport into temperature and salinity space, or $M(T,S)$. Specific focus was maintained on identifying characteristic features of the ocean circulation and the major water masses within the control run scenario. $M(T,S)$ maps were generated for 5 specific latitudes, namely 34°S, 20°S, 0°N, 24°N, and 40°N, with maps belonging to both the global integral, and the Atlantic basin. Figure 8.2 shows these $M(T,S)$ maps at 24°N for CHIME and HadCM3.

Near surface waters are characterised as having a larger spread in water mass locations
on the M(T,S) maps, consistent with wind-induced circulation that tended to mix water mass properties over the first 1000m. Deeper within the ocean interior, water masses tended to cluster around specific isopycnal surfaces with a smaller spread in location in T-S space. Atlantic water masses tended to be saltier compared to the Indo-Pacific region, resulting in a clear separation of water mass features for the two basins, particularly for the NH. Further southwards the characteristic features of the two basins tended to merge together, overlapping almost completely at 34°S, where the close proximity of the Southern Ocean mixed the major waters from all major ocean basins together.

A comparison of M(T,S) maps between CHIME and HadCM3 for identical latitudes revealed very similar features, with the major water masses sharing similar T-S spatial coordinates and T-S curve distributions. A key change when switching between the different vertical coordinate systems is that whilst CHIME’s water masses laid on clearly defined isopycnal surfaces, HadCM3 had much diffuse properties, with water masses spreading across isopycnal surfaces. This is to be expected since CHIME’s vertical coordinate is already density. Another
key difference is the relative ease in identifying the major water masses in T-S space. Megann et al. (2010) showed that CHIME had a superior representation of internal water masses and their spatial preservation with time, particularly the AAIW as compared with HadCM3. This has been found to extend to other water masses such as NADW and AABW with their identification being better defined in CHIME, primarily owing to the better preservation of water mass properties that results in the less diffusive nature of water mass locations in T-S space.

In Chapter 5 we developed a new approach to analysing the coupling between heat and freshwater transport at extra-tropical latitudes. We built on theory derived by Stommel and Csanady (1980) that stated the ratio of ocean heat transport $H_o$ to freshwater transport $F$ across a fixed latitude circle, was proportional to the ratio of the temperature and salinity gradient (difference between integrated poleward and equatorward moving values) $\Delta T/\Delta S$, or more commonly known as the T/S curve:

$$\frac{H_o}{F} \approx c_o S_o \frac{\Delta T}{\Delta S} \quad (8.1)$$

where $c_o$ is the specific heat capacity of seawater and $S_o$ is the zonally mean salinity. Since in steady state, the magnitude of ocean freshwater transport in the ocean is of the same magnitude and opposite direction in the atmosphere, the atmospheric heat transport was expressed as function of freshwater transport and the quantity $\gamma = DSE/L$, defined by the ratio of dry static energy DSE to latent heat transport $L$ ie. $H_a = l_v F (1 + \gamma)$, where $l_v$ is the enthalpy of vaporisation of water. $\gamma$ was computed directly from calculations of freshwater transport and atmospheric heat transport, however a further simplification proved that $\gamma$ could be approximated simply as a function of zonal mean SST ie. $\gamma \approx \gamma (SST)$. Combining all the above, generated,

$$\frac{H_o}{H_a} = \frac{c_o S_o}{l_v} \frac{1}{1 + \gamma (T)} \frac{\Delta T}{\Delta S} \quad (8.2)$$

Hence the ratio of ocean to atmospheric heat transport is proportional only to the temperature and salinity variations in the ocean. This relationship was tested using the HadCM3 model in a control climate scenario (a different control to the one previously used), with the results shown in Fig. 8.3. Agreement of the theory to direct model calculations of $H_o/H_a$ was between 80-90% for the NH, depending on the choice of calculation for $\gamma$, and less than 20% for the SH. The discrepancy between inter-hemisphere accuracy was linked to the unavailability of ocean bolus velocities in our data set at the time, with only the resolved advective transport
used. Transport belonging to bolus velocities were found to be particularly high for the SH region compared to the NH, leading to the difference in accuracy already stated.

Figure 8.3: The ratio $H_o/H_a$ in HadCM (solid) and predicted from (5.17) using $\gamma_F$ (dashed). Plots in black (grey) represent the Northern (Southern) Hemisphere. The dotted lines corresponds to an additional calculation in which $\gamma = \gamma_T$ is predicted rather than calculated in the model. Adapted from Figure 5.6.

A form of compensation between $\Delta T/\Delta S$ and $\gamma$ was observed between the two hemispheres, suggesting that different ocean mechanisms to the ocean circulation eg. large-scale overturning, vs eddy-driven circulations, can contribute to different values.

The theory does have its limitations: firstly it is only valid for extra-tropical latitudes where latent heat transport is poleward. Secondly, the relationship breaks down as surface temperatures reach freezing point. At this stage, ice cover over the ocean separates the necessary coupling of freshwater and salinity needed between the ocean and atmosphere. Thirdly, the model has to be integrated over a suitable period of time such that there the magnitude of freshwater transport in the ocean and atmosphere are in balance with each other. Generally, any perturbation in the atmosphere would take centuries to fully translate any effects throughout the entire ocean water column. In the views of present climate change, the theory is best suited to control climates with fixed $CO_2$ values instead of transient scenarios.

The theory was not designed to offer a better way at computing the ratio of ocean to atmospheric heat transport, since satellite measurements already offer a far superior methodology (see Trenberth and Caron, 2001). What it does offer is a new and highly simplified framework to understanding this constraint on the system ($H_o/H_a < 1$ at extra-tropical latitudes) that has never previously been derived. Additionally, any research into the paleo-reconstruction of heat transport would need only to reconstruct the past ocean profiles for temperature and
salinity.

In Chapters 6 and 7 we looked at the climatic responses of heat and freshwater transport to increasing $CO_2$ using CHIME and the EPcm model only. A 200 year transient climate scenario was used that increased $CO_2$ concentration by 1% per year from pre-industrial levels starting at year 60, and ending at $4 \times CO_2$ concentration. For CHIME 4 snap-shot periods were chosen centred at $CO_2$ concentrations: $1.4 \times CO_2$, $2.0 \times CO_2$, $3.0 \times CO_2$, and $3.8 \times CO_2$ (4), each period being 10 years in length. We defined anomalous scenarios as the change between the control climate and the 4 different transient scenarios.

Chapter 6 examined the main robust responses of CHIME’s climate to an increase in $CO_2$ concentration. These were found to include:

- increase in global mean surface temperatures, with NH high latitudes warming faster than for the equatorial regions, leading to a reduction in the equator-to-pole temperature gradient
- SH cold anomalies were found in the Southern Ocean region leading to an enhancement of the meridional surface temperature gradient
- enhancement of the pattern of evaporation and precipitation, with the sub-tropics and extra-tropical latitudes getting wetter, and the tropics becoming drier
- increase in column integrated global mean values of specific humidity at the Clausius Clapeyron rate of $7.5\% K^{-1}$
- increase in global mean precipitation at a rate of $1\% K^{-1}$
- northward shift of the tropical Pacific ITCZ, though no overall shift for the ITCZ’s global position.
- increase in SSS over the Atlantic and Arctic regions, with a corresponding freshening of the Indo-Pacific and Southern Ocean regions

all of which were consistent with other model predictions from the latest IPCC report (Solomon et al., 2007; Held and Soden, 2006; Lorenz and DeWeaver, 2007). Inspiration was
drawn from the last point mentioned about changes to SSS. The positive salinification of the Atlantic and corresponding freshening of the Pacific imply a net freshwater transport from the former to the latter region. Just as in Chapter 3, we hypothesised that this freshwater would be transported through the Panama Isthmus region via the Easterly Trade winds. This would be a net freshwater transport out of the Atlantic region, with any return flow of moisture via the Westerly Winds at mid-latitudes being blocked by orographic features (see Fig. 8.4).

Figure 8.4: SSS anomaly at $4 \times CO_2$ concentration. Black arrows indicate atmospheric flow of moisture, whilst brown rectangles represent orographic features to block flow. Equatorial Eastern Pacific region defined by dashed box region with red arrows represent net direction of anomalous moisture transport into and out of the region. Adapted from Figure 3.19.

Although this view was not conclusive when comparing models HadCM3 and CHIME under a control climate scenario, we employed this view again since the perturbation to the system was now sufficiently large, spanning a quadrupling of $CO_2$ to account for any statistically significant response to atmospheric freshwater transport. A region in the Eastern Equatorial Pacific (EEP) region was chosen for analysis, bounding the coordinates $200 - 260^\circ E$ by $0 - 20^\circ N$. Calculations of net freshwater transport into the region to satisfy the model changes to SSS were found to be consistent with direct calculations of net evaporation minus precipitation, showing that indeed changes to SSS were forced by net changes of freshwater at the surface. To understand the sources of this atmospheric freshwater transport into the region, direct calculations through all 4 sides were compiled. The results showed that the main sources originated from the east and south sides, indicating that there is indeed an increase of freshwater transport from the Atlantic into the Pacific region, however the input from the south side was consistent with a regional northward shift of the ITCZ. Changes to the Pacific freshwater
budget thus displayed both regional and inter-basin mechanisms.

Similarly for Chapter 6, we examined how meridional HT changed with increasing $CO_2$ (see Fig. 8.5). The main results pertaining to the ocean and atmospheric media are summarised below:

- ocean (atmospheric) HT decreases (increases) by similar magnitudes so as to conserve total HT, in line with the ideas of Bjerknes Compensation
- differences between the ocean and atmosphere are statistically significant (at the 1 standard deviation level) for $CO_2$ greater or equalling to triple pre-industrial values
- compensation is strongest at extra-tropical latitudes greater than $40^\circ$ for both hemispheres
- latter latitude penetrates down to almost $20^\circ S$ for the SH

The decrease in the ocean mass transport was examined by decomposing the latter into parts belonging to meridional mass transport $M_o$ and corresponding temperature gradient $\Delta T$. Particular attention was paid to the Atlantic basin and its meridional overturning circulation (MOC) since the region is the main carrier of ocean HT northward of $40^\circircN$. The main results are summarised:

- decrease in ocean HT was better correlated to decreases in $\Delta T$ over both hemispheres than to changes in $M_o$
- decrease in $\Delta T$ would suggest a warming of equatorward moving water at a rate faster than its poleward moving counterpart
- examination of the Atlantic basin showed ocean interior warming at a depth matching the southward flow of NADW
- MOC pathway suggests the interior warming originates at high NH latitudes

The first bullet point described above has been found to be the complete opposite to real world observations where variability of ocean HT is primarily controlled by volume transport (Wunsch and Heimbach, 2006). It is worth to note that the ocean HT variability used annual mean data only (10 years for each transient period). We suspect though that if monthly mean data was used instead, ocean HT variability would be coupled stronger to mass transport since
Figure 8.5: Anomalous differences in meridional heat transport for the total (blue), atmosphere (red), and ocean (blue) heat components for the 4 transient scenarios. Error bars signify a ± 1 s.d. for the latitudinal dependent time series.
for instance, the variability of the MOC would be controlled by the deep convection of water that only happens during winter months (REFERENCE).

Similarly, the treatment of atmospheric HT was decomposed into parts belonging to that for latent and DSE transport (see Fig. 8.6). The main results of our analysis are as follows:

- increase in atmospheric HT was a result of an enhancement of latent HT, even though DSE was found to decrease
- increase in atmospheric HT largest over extra-tropical latitudes greater than 40°N
- differences between latent and DSE were observed to be significant only for concentrations greater or equal to a doubling of pre-industrial CO₂ levels
- at these latitudes, changes to the latter two components are similar to Bjerknes Compensation with net changes to atmospheric HT acting as a constraint, similar to total HT constraining the activities of the ocean and atmosphere
- at 4 × CO₂ concentration the largest changes to the internal composition of atmospheric HT shift from the extra-tropics to the tropical SH region, with an enhancement of both latent and DSE transport by approximately 1.5 PW
- analysis of the SH Hadley Cell found that circulation strength increased by 15% after an almost quadrupling of CO₂, with values of low level specific humidity increasing at the Clausius-Clapeyron rate. The enhancement of both components was enough to explain the increase in latent heat transport over the region

Explicit calculations for changes to DSE proved inconclusive. Since DSE is comprised of the sensible and gravitational potential energy fluxes, calculations for both these sub-components were carried out. An examination of the maximum height of the Hadley Cell and the height at which maximum overturning was located revealed no shift in both positions, inferring no change in gravitational potential energy. An area weighted calculation of the potential temperature difference between northward and southward moving air parcels revealed that there was a slight increase in temperature difference, however the latter increase and the enhanced Hadley Circulation strength was not enough to take into account the change in DSE after 4 × CO₂. The issue still remains unsolved. The main conclusions we would draw from changes to atmospheric HT would be that a). it increases, b). the increase is driven by an enhanced latent heat transport, and c). regions of significant change shifts from the extra-tropics towards the tropics for very high values of CO₂ concentration.
Figure 8.6: Similar to Fig. 6.8 but for differences in meridional heat transport for the atmosphere (blue), latent heat (red), and dry static energy (blue) heat components.
In Chapter 7 we examined changes to the ocean circulation with increasing $CO_2$. A combination of techniques were involved ranging from M(T,S) analysis to plotting mass transport as a function of density $\sigma_t$.

A key area of study involved the evolution of the Atlantic MOC, taken as the volume streamfunction maximum over the entire basin. The time series evolution is shown in figure 8.7 with key results from our analysis summarised below:

- at the onset of $CO_2$ increase, MOC strength slightly increases for 40 years to a maximum of about 21 Sv at year 100, before decreasing in strength for the next 100, reaching 16 Sv at year 200
- changes to MOC strength were found to be highly anti-correlated to variations in net surface heat flux loss for the deep convection regions of the Labrador Sea and locations off the coast of Greenland
- a cross-correlation between MOC strength and net surface heat flux loss was found to be highest when the latter led by 5 years

Figure 8.7: The maximum strength of the AMOC is shown by the red line (Sv) within the transient $CO_2$ scenario forced at at% increase per year. Area weighted net surface heat flux over the Deep Convection Region for February months is represented by the blue line. The latter has been rescaled by the multiple -9.5 with a 10 year smoothing filter over the time series to remove some variability. Adapted from fig. 7.8.
A puzzling problem with CHIME was the observed positive salification of SSS in the Atlantic basin with increased $CO_2$ as shown in Fig. 8.4, even though the MOC weakened. The numerical experiments of Manabe and Stouffer (1988) found two stable equilibria of the Atlantic basin, one having an active thermocline circulation and the other without. For the former configuration they found that the active thermohaline circulation advected high salinity waters from the Tropics towards the poles, whilst in the latter, the near disappearance of a thermocline increased the residence time of near surface waters to the effects of net precipitation at high latitudes. There was thus this link between high SSS values and MOC strength in the Atlantic. We however explain that the discrepancy in CHIME’s anomalous SSS signature can be congruent with a weakening MOC:

- propagation of very high SSS anomalies (> 0.6 psu) was observed to occur at about year 110, 10 years after the start of the decreasing MOC
- generation of high SSS anomalies in the NH Tropics is enhanced owing to increased net evaporation over the region
- slowing of Atlantic MOC increases the residence time of near surface waters to the net evaporative Tropics, sustaining positive SSS anomalies over a longer distance

Although there is a northward propagation of high SSS values in the Atlantic acting to decrease surface buoyancy, the effects of surface warming from anthropogenic forcing acts to counter this effect such that the northward advection of water becomes lighter than that for the control run with pre-industrial $CO_2$ values. Any stabilisation mechanism of the MOC from the advection of high SSS (Latif et al., 2000; Thorpe et al., 2001) would most likely not happen within the CHIME model.

For the Epem model, changes to its climate with increasing $CO_2$ yielded similar trends to that of CHIME i.e. increasing surface temperature, Clausius-Clapeyron scaling, and an enhanced hydrological cycle. The main difference in response was that latent heat transport decreased with $CO_2$ concentration; a surprising result given the aqua planet model setup. Figure 8.8C illustrates this feature with the black square (latent heat transport at double $CO_2$ concentration) being lower in value than the black circle (latent heat transport at pre-industrial values). This decrease was found to be primarily coupled to the decrease in atmospheric circulation strength $Ψ_A$ being stronger than the increase in the model specific humidity gradient $Δq$. By varying the model water vapour emissivity parameter $γ$ by multiples of the default value (0.1 blue, 0.2 cyan, 0.5 green, 1.0 (default) black, 2.0 yellow, 5.0 red, and 10.0 magenta)
it was found that the response of latent heat transport to increasing CO₂ would increase with CO₂ for multiple values of γ less than or equal to 0.6 times the default value.

Figure 8.8: Responses of A), surface temperature gradient, B), specific humidity gradient, C), latent heat transport, and D), atmospheric mass transport, to variations in water vapour emissivity parameter γ. Coloured points refer a multiple of the default parameter value: 0.1 blue, 0.2 cyan, 0.5 green, 1.0 (default) black, 2.0 yellow, 5.0 red, and 10.0 magenta. Blue (red) curves correspond to climate scenarios at pre-industrial (double) CO₂ values. Identical to fig. 3.7.

Lastly we analysed how the ocean circulation changed through the examination of M(T,S) maps under an increasing CO₂ environment in CHIME. Using the same latitudinal cross-sections mentioned for the control climate, the results obtained helped to verify the previous analysis already mentioned.

- the majority of water masses located on M(T,S) maps shift wards, consistent with a warming of the entire ocean
- water masses for the Atlantic (Indo-Pacific) tended to shift towards the right (left) in T-S space, inferring a salinification (freshening) of waters

By plotting meridional mass transport as a function of density it was possible to identify many of the major water masses, particularly those exclusive to the Atlantic basin. Of the three major water masses in the ocean i.e. NADW, AABW and AAIW, our analysis of how their flow rates changed with increasing CO₂ found that:

- southward transport of NADW decreases at a similar rate to the reduction in strength of Atlantic MOC
Figure 8.9: T-S map at 40°N for the global zonal ocean cross-section for the control climate (blue) and at 4 × CO₂ (red). Water masses belonging to the Atlantic (Indo-Pacific) are represented by all water masses greater (less) than 35 psu. Red arrows indicate general direction of water mass distribution for increasing CO₂. Adapted from Fig. 7.6A.

- northward propagation of AABW decreases, affected by a reduction in ice cover fraction and a decrease in net surface heat loss off southwards of 57°S
- northward flow of AAIW increases, partially a result of an enhanced northward Ekman transport over the Southern Ocean from an increase in zonal wind stress

Figure 8.10: Left figure: meridional mass transport as a function of density at 20°S across the globe (left column), the Atlantic basin (middle column) and the Indo-Pacific basin (right column). Blue (red) distributions indicate results for the control climate (4 × CO₂ scenario. Right figure: corresponding M(T,S) map at 20°S. Red (blue) colours represent northward (southward) flowing water masses. Adapted from Fig. 7.10D.

8.2 Future Work

Due to the broadness in subject of this thesis it has not been possible to address all issues within the allotted time period. More often than once, searching for an answer to a problem...
often raised more questions than answering it. This was never found to be a negative issue, but hinted to the possibility of future research. Below is a list of these issues

1. In Chapter 3 we compared the meridional HT between CHIME and HadCM3 within a control run scenario. Although the decrease in the NH Hadley Circulation strength was inferred to be due to the reduced equator-to-mid-latitude temperature gradient across the region, this is not explicitly proved. To fully prove this assumption a numerical experiment could be run with the CHIME model forced with only the SST temperature difference between the two models. If the model shows a reverse NH Hadley Cell then the conclusion would be verified.

2. In Chapter 5 we examined new theory that linked how the ratio of ocean to atmosphere HT could be calculated from the T/S curve and the atmospheric parameter $\gamma$. Given the formula’s simplicity it would be interesting to see how the relationship between the latter two variables changes for climate scenarios with different $CO_2$ value concentrations. Since the theory assumes that the ocean and atmospheric freshwater transport are in steady state, only climate states with stable $CO_2$ values could be analysed. Though this may not be relevant to a transient $CO_2$ environment such as the one used in this thesis, it can offer interesting insights into paleo-climatic times such as the warm Eocene period and the cold Last glacial Maximum where $CO_2$ were highly different but stable to variations in comparison to present day observations.

8.3 What Have We Learnt?

The study of climate physics still has many challenges ahead, not least by the spread in GCM uncertainty arising from differing resolution, coordinate choice, and parameterisations. Within an ocean model, the difference between z-level and $\sigma_t$-level vertical coordinate choice are found to have significant changes to the overall climate, especially within the atmospheric decomposition of meridional heat transport into dry static energy and latent heat. Given that these discrepancies are observed within a control climate scenario, it is quite worrying how these differences would evolve when integrated under increasing anthropogenic forcing. On the positive side, present day GCMs do model the climate system with a high degree of fidelity, if only on the regional and on shorter time scales, such as daily weather forecasts. Though climate predictions for the next century still have a large uncertainty attached, the general trends are at least pointing in the same direction to patterns predicted by Physics. Within
the analysis of the GCM CHIME, these included the overall increase in global mean surface temperatures, the enhancement of the hydrological cycle, the increase in global mean column integrated water vapour content, and the reduction (enhancement) in ocean (atmospheric) heat transport, all congruent with past model estimates.

Parameterisations are a powerful tool. A simple theory that couples the ratio of Dry Static Energy to latent heat (freshwater) transport in the atmosphere simply from the surface temperature was proven to be consistent in a GCM. This parameterisation has the added advantage of being a new diagnostic for any model to either verify its physics, or as a computational shortcut for the calculation of heat transport. Additionally, there is still no proof that further increasing the resolution of a GCM will lead to a better representation of the climate. A model with a coarser resolution, but with superior parameterisations has a chance of being the better model. Given this, the author feels that there will be a push towards models with higher resolutions in the future, made possible due to the constant improvements in computational power.

Changes to ocean and atmospheric heat transport under increasing CO$_2$ have been well documented to behave according to the ideas of Bjerknes Compensation. In our examination of CHIME however it was found that significant differences are only achieved at concentrations greater than a tripling of original CO$_2$ concentration due to the large amount of model internal variability. At present projections this level of CO$_2$ would be reached after approximately 100 years. Given our present data records, the verification of such compensation would be impossible to make for many years to come.

Finally, the study of Climate and Climate Change, will always be an incomplete science, reliant on ever increasing computer power that makes yesterday’s predictions obsolete, to the highly variable projections of GHG emissions. There will always be a refinement to the latest climate predictions with the addition of a new parameterisation or model, however whether the predictions are good enough, especially for the general public, is a question extremely in need of investigation.
A.1 The Material Derivative

For a fluid parcel at a fixed location $x$, we can define its time dependence as $\gamma = \gamma(x, t)$. If the parcel is moving, its displacement is time dependent, resulting in $\gamma = \gamma(x(t), t)$.

The time rate of change of the parcel can be expressed as

$$\frac{d\gamma}{dt} = \frac{\partial \gamma}{\partial t} + \frac{\partial \gamma}{\partial x} \frac{dx}{dt} + \frac{\partial \gamma}{\partial y} \frac{dy}{dt} + \frac{\partial \gamma}{\partial z} \frac{dz}{dt} \quad (A.1)$$

$$\equiv \frac{\partial \gamma}{\partial t} + \frac{d\bar{x}}{dt} \cdot \nabla \gamma \quad (A.2)$$

Since $\frac{d\bar{x}}{dt} = \bar{u}$,

$$\frac{D\gamma}{Dt} = \frac{\partial \gamma}{\partial t} + u \frac{\partial \gamma}{\partial x} + v \frac{\partial \gamma}{\partial y} + w \frac{\partial \gamma}{\partial z} \quad (A.3)$$

$$\frac{D\gamma}{Dt} = \frac{\partial \gamma}{\partial t} + \bar{u} \cdot \gamma \quad (A.4)$$

This is called the Material Derivative.
APPENDIX B

Appendix for Chapter 5

B.1 Derivation of γ_{sst} - Eq. (5.21)

To derive equation (5.21) we limit ourselves to mid-latitudes, where the main mechanism of latent heat transport is through atmospheric eddies,

\[ l_v F \approx l_v \int \frac{v' q'}{g} dP \]  \hspace{1cm} (B.1)

At fixed relative humidity RH, greater air temperature \( T' \) correspond to an increase in the specific humidity \( q' \),

\[ q' = \left( \frac{dq_{\text{sat}}}{dT} \right) RH \times T' \]  \hspace{1cm} (B.2)

where we have rewritten \( q = RH \times q_{\text{sat}} \), with \( q_{\text{sat}} \) being equal to the saturation specific humidity. Relative humidity is set to 0.8 which is close to lower tropospheric values (Pierrehumbert, 2002). By multiplying the above equation by \( v' \), using our approximation for latent heat transport and our expression for \( \gamma \) in Eq. (5.11), we can derive a new expression of \( \gamma \) such that

\[ \gamma = \frac{c_p}{l_v RH} \left( \frac{dq_{\text{sat}}}{dT} \right)^{-1} \]  \hspace{1cm} (B.3)

This can be developed further by using the Clausius-Clapeyron expression linking humidity changes to temperature changes (Hartmann, 1994)

\[ \frac{dq_{\text{sat}}}{dT} = \frac{l_v}{RT_s q_{\text{sat}}} \]  \hspace{1cm} (B.4)
where \( R = 287 J K g^{-1} K^{-1} \) is the Gas Constant for water vapour and \( T_s \) is the surface temperature. By combining the latter expression with (B.3), we readily obtain,

\[
\gamma_{T_s} = \frac{c_p R}{l_f^2 R H} \frac{T_s^2}{q_{sat}(T_s)}
\]  

(B.5)

The latitudinal distribution of \( \gamma_{T_s} \) for both hemispheres can be seen in Fig.5.5b (dashed lines) in which we have used zonal average values for \( T_s \). The model values i.e. \( \gamma_F \) (solid lines) shows similar values of approximately 2 at middle latitudes (40°N), which is of similar magnitude to NCEP observations of \( \gamma \approx 1.4 \) (Pierrehumbert, 2002). The increase in magnitude with latitude is due to latent heat transport decreasing at a faster rate as compared to dry static energy, since at colder air temperatures the atmosphere’s ability to store moisture is greatly reduced. The agreement between \( \gamma_{T_s} \) and \( \gamma_F \) is good in the Northern Hemisphere but, as was found for the ratio \( H_o/H_a \), poor in the Southern Hemisphere.
C.1 Fixed and Transient CO₂ scenarios of EPcm

The two box model allows for varying CO₂ concentrations in the atmosphere to modulate the net radiation into each region. This has feedback into the climate whereby quantities vary slightly to the changes in CO₂ concentration. One can therefore use this model to study climate predictions for different climate scenarios. Within this section we analyse the changes to the climate system using a variety of CO₂ emission scenarios, both fixed and transient in time.

For the fixed CO₂ scenarios we take multiples of the control run concentration of 280 ppm: 1.0, 2.0, 3.0, and 4.0 times, and integrate each in time for 300 years. A transient CO₂ run is also used whereby the concentration increases from 280 ppm by 1% per year up to 4 times the control concentration, or 1120 ppm. This value is reached after 140 years of integration time whereby afterwards the model is held at its present CO₂ concentration and allowed to stabilise for another 160 years. Thus both fixed and transient CO₂ scenarios are run for 300 years each (see Table C.1 1st row). We note that this is also the time taken for ocean freshwater fluxes to balance out that within the atmosphere. As a point of 2×, 3×, and 4× CO₂ concentrations happen at year 70, 110 and 140 respectively for the transient simulations.

For fixed CO₂ scenarios, variables are computed as the final value at the end of the 300 year integration run and represented as symbols. The exception to this is variables that
Appendix C

Table C.1: Graphical representation of all scenarios as a function of both varying CO\textsubscript{2} concentration and fixed Ψ. The numbers represent the multiple CO\textsubscript{2} concentration with 280ppm set as the default climate; ‘T’ represents the transient CO\textsubscript{2} scenario with concentrations increasing by 1% per year until 4× default concentration at year 140, then sustained at this level for the next 160 years. Fixed Ψ scenario are represented by the symbols: ∗ (no fixed Ψ, default settings), ○ (fixed Ψ\textsubscript{O}), Δ (fixed Ψ\textsubscript{A}), and □ (fixed Ψ\textsubscript{O} and Ψ\textsubscript{A}). Transient runs for each different Ψ scenario are represented by curves. In total, there are 20 different scenarios for analysis.

have random stochastic noise added to its integration run whereby its final value fluctuates i.e. global net TOA radiation. For these variables a mean value of the last 365 days of the integration is taken to remove this noise. All symbols are aligned to the years 0, 70, 110 and 140, coinciding with the scenario fixed CO\textsubscript{2} of the transient run (see Fig. C.5a). Transient runs are represented by curves since it is a varying time evolution of the variable. Although both fixed and transient CO\textsubscript{2} scenarios are integrated up to 300 years each, only the first 200 years are shown to magnify the details of symbol scenarios. For the majority of variables the last 100 years of transient integration is slow varying, thus omitting this period does not significantly impede the interpretation of results.

Although variables change under varying amounts of CO\textsubscript{2} forcing it is sometimes difficult to tell whether these changes have a stronger coupling to changes in either the ocean (Ψ\textsubscript{O}) or atmosphere (Ψ\textsubscript{A}). Given the simplistic architecture of EPcm it is possible to decouple the effects of both media by method of holding one constant whilst letting the other vary with time. These fixed Ψ scenarios include fixed Ψ\textsubscript{O} (circles, dashed line), fixed Ψ\textsubscript{A} (triangles, dotted line), fixed Ψ\textsubscript{A} and Ψ\textsubscript{O} combined (squares, dashed-dotted line), and the default simulation with no variables held (stars, solid line); a fixed freshwater transport simulation is also described but not included within the plots as the changes within this scenario are very small and therefore would ‘crowd’ the details of the other scenarios. A schematic of all the different scenarios under different levels of CO\textsubscript{2} concentration and under fixed Ψ conditions is shown in Table C.1.
For brevity, changes to variables described for both fixed and transient $CO_2$ scenarios are all relative to its equilibrium control run value.

**C.2 Fixed $CO_2$ scenarios - default $\Psi$ conditions**

For clarity, this subsection will only refer to the star symbols on plots.

The Thermocline temperature increases in both regions, however after 300 years (except for the control run) the thermocline layer still not fully reached its equilibrium temperature i.e. $T_{O1} \neq T_{O2} \neq T_{S2}$, with the difference in values between $T_{O1}$ and $T_{O2}$ becoming larger with $CO_2$ concentration (Fig. C.1b). One can however infer the eventual equilibrium temperature of the thermocline layer by comparing it to the temperature of $T_{S2}$, since this region’s temperature properties are conserved and advected into the thermocline layer. There is a successive increase in surface temperatures within each region, with global temperatures increasing by $2.69^\circ C$ after $2 \times CO_2$ concentration. The change in global mean temperature is largest for smaller multiples of concentration, with differences between adjacent multiple states becoming successively smaller (Fig. C.1C).

Ocean heat content (Fig. C.2b) mirrors the trends in ocean temperature as waters become warmer due to the increase in global surface temperatures. This heating effect, especially of mixed layer waters is non-uniform, with the extra-tropics warming at a faster rate than the tropics; an effect known in literature as the *Polar Amplification* effect that leads to a reduction in the equator to pole temperature gradient (Fig. C.1d).

Global mean net TOA radiation increases with $CO_2$ concentration such that even after 300 years of integration with fixed $4 \times CO_2$, there is a positive net value of approximately $0.5 W m^{-2}$ (Fig. C.2a). Normally for the control climate, the system reaches equilibrium within 300 years of integration, resulting in a net radiation at the TOA being close to zero. The net absorption of surface heat flux at the surface in the Tropics would directly balance out the net emittance of radiation within the Extra-Tropics. With increasing $CO_2$ the Tropics (Extra-Tropics) receives (emits) less radiation. The decrease in the latter region is smaller due to the initiation of evaporative cooling at high latitudes (Fig. C.2d, blue curve) as the climate warms. This in turn leads to the smaller decreases in extra-tropical net surface radiation that yields the net positive imbalance of radiation at the TOA. As stated above for the control climate, these imbalances ultimately return to zero when the climate system is in equilibrium.

For the large increases in $CO_2$ concentration, 300 years of integration time is not enough, the
ocean being the dominant system to reach equilibrium with its surroundings. A crude linear extrapolation of net TOA energy flux converging to zero with time would suggest a waiting of approximately 1000 years, consistent with oceanic timescales.

Unlike temperature, the salinity evolution of the ocean remains fairly consistent throughout the different $CO_2$ concentrations, with thermocline layer values reaching equilibrium within 300 years of integration time (Fig. C.3a, b). The salinity gradient increases due to the enhancement of evaporation over the Tropics and precipitation within the Extra-Tropics (Fig. C.3d).

There is an increase in the salinity gradient due to the enhanced evaporation (precipitation) within the box 1 (box 2).

Since atmospheric circulation strength is directly proportional to the temperature gradient, there is a decrease in $\Psi_A$ (and therefore $\Psi_O$) of $-4.3\%$ per K warming between control and $2 \times CO_2$ runs (Fig. C.4c). As a result the atmospheric heat transport decreases by $-5.9\%$ per K warming (Fig. C.4a). The oceanic heat transport decreases by $-8.1\%$ per K warming due to polar amplification effects and a reduction in the oceanic circulation strength. The total heat transport of the system therefore decreases by $-0.77PW$ between $2 \times CO_2$ and control run scenarios (Fig. C.4b).

We note that in the real world, changes in oceanic circulation strength lag that of the atmosphere due to both wind and buoyancy driven forces in the ocean. Although the model does contain salinity it is an inactive tracer and therefore does not contribute an buoyancy driven effects to $\Psi_O$. The oceanic circulation strength is tightly coupled to the atmosphere via $\Psi_O/\Psi_A = 0.1$ and is therefore instantaneous. Although this approximation may work well with fixed $CO_2$ runs (provided that the integration time is long enough for the system to reach equilibrium), this is a poor approximation when analysing transient $CO_2$ scenarios since the ocean adjustment time to any atmospheric perturbations in $CO_2$ are on the order of centuries.

Specific humidity contrast between low and high latitudes increases (Fig. C.5b). This amounts to a 6.4% increase in low level moisture predicted by Clausius-Clapeyron scaling. Specific humidity is a function of temperature i.e. $q = q(T)$; even though the temperature contrast decreases (Fig. C.5c), changes in low latitude values of $q$ are still larger than at high latitudes due to the non-linear relationship between specific humidity and temperature.

If we decompose the atmospheric heat transport into latent heat transport $L$ and dry static energy DSE $(H_A - L)$, much of the decrease in $H_A$ originates from the reduction in DSE (Fig. C.5d); changes in latent heat transport plays a minor role. To understand the different
reductions in $H_A$ we look at each variable’s dependence on the temperature gradient since this is the main controlling parameter of meridional heat transport. From simple mathematical manipulation we can show that $\text{DSE} \propto (\Delta T_A)^2$, whilst $L \propto \Delta T_A \Delta q$. Since $\Delta T_A$ decreases with $CO_2$ concentration, the change in DSE is magnified by its squared dependence on the temperature gradient. In comparison with L, since $\Delta q$ increases, this partially compensates the reduction in the temperature gradient allowing for a more insensitive Latent heat transport to variations in $CO_2$.

Freshwater transport decreases with increasing $CO_2$ concentration, however its rate is less than that of Clausius-Clapeyron, at $-4.3\%$.

The ratio of ocean to atmospheric heat transport increases slightly due to the decrease in $H_A$ being larger than that of $H_O$ (Fig. C.6b). Given this increase the magnitude of this ratio still remains fairly constant at $\approx 0.4$ even at $4 \times CO_2$ concentration. Since both $H_O$ and $H_A$ decrease with increasing global mean surface temperatures, the model does not exhibit the classical signatures of Bjerknes Compensation, whereby changes in either heat transport component is directly balanced by an equal and opposite change with its counterpart.

The ratio of DSE to L in the atmosphere, $\gamma$ decreases due predominantly to changes in DSE (Fig. C.6c), whilst the slope of the T/S line, or $\Gamma = \Delta T/\Delta S$ also decreases (Fig. C.6d). The changes however in $\Gamma$ are smaller than that of $\gamma$. Although not included within this chapter, there is theory linking the ratio of ocean to atmospheric heat transport as a function of only these two parameters. We briefly conclude that for EPcm, changes to $H_O/H_A$ are coupled stronger to the variations in the atmospheric parameter $\gamma$ rather than in the oceanic $\Gamma$.

### C.2.1 Transient $CO_2$ scenarios

For the transient $CO_2$ scenarios many of the aforementioned variables have almost instantaneous reactions to the changing $CO_2$ content, with little lag time for variables that are predominantly coupled to atmospheric processes i.e. atmospheric temperature, specific humidity, circulation strength etc. Thus at year 140 (transition between increasing $CO_2$ concentrations and $CO_2$ termination) the variable’s time evolutionary progress, in many cases, instantaneously stabilises under constant $CO_2$ for the next 160 years, leading to the visible ‘kink’ in many of the curves. Variables that are coupled to the deep ocean (except the inactive model tracer, salinity ) i.e. ocean temperature and heat content, have a relatively larger lag period between transient and $CO_2$ runs since the communication of changes in the atmosphere’s $CO_2$ is limited by the strength of the ocean overturning system advecting ‘information’ into the deep
Appendix C

Figure C.1: Model temperature evolution with time: Box 1 tropical region for mixed layer (red) and thermocline (magenta) a). Box 2 extra-tropical region for mixed layer (blue), thermocline (cyan), with inserted tropical thermocline layer (magenta) for comparison b). Global mean surface temperature c). and surface equator to pole temperature gradient d). CO$_2$ concentrations of fixed and transient evolutions are represented by symbols and curves respectively. Fixed variable scenarios are identified for fixed $\Psi_O$ (circles, dashed line), fixed $\Psi_A$ (triangles, dotted line), fixed $\Psi_O$ and $\Psi_A$ (squares, dashed-dotted line), and no fixed variables (stars, solid line).

The long lag time for the ocean reaching equilibrium with its surroundings is primarily responsible for the net radiation imbalance at the TOA under increased CO$_2$ forcing.

C.3 Appendix III: The sensitivity of the EPcm model to fixed $\Psi$ under different CO$_2$ environments

The previous section dealt with how variables changed under different CO$_2$ forcing scenarios. We have seen how different climatic 'observables' can change under different CO$_2$ forcing. However, since the changes in different variables are usually dependent on changes in more than one other variable e.g $F = \Psi_A \Delta q$, it is sometimes hard to decouple these changes with the aim of analysing the main drivers of the change in the system. For this reason we have re-run the above scenarios but with certain variables set at fixed values such that they are independent of CO$_2$ concentration. These scenarios are: fixed $\Psi_o$, fixed $\Psi_a$, fixed $F$, and fixed $\Psi_o$ and $\Psi_a$. We have picked these specific variables as these are essentially the main drivers of meridional heat transport. By effectively decoupling the ocean and atmosphere we can probe
Figure C.2: Evolution of net radiative fluxes at the TOA and Surface as a function of time: Global net TOA radiation a). Ocean heat content b). Net surface heat flux c) and evaporative cooling d) for boxes 1 and 2. The sign convention is such that positive (negative) values represent net emittance (absorption) of radiation at the interface. Curves for subplots a), b), and d) have been smoothed using a 365 day filter to remove the stochastic forcing signal. Fixed CO$_2$ symbols for the mentioned subplots are the mean value for the last 1 year of the time series.

deep into the understandings of how different the climate can alter in these scenarios and therefore the importance of a coupled system.

We again highlight that the changes to variables under fixed conditions are in comparison to the default scenarios.

C.3.1 Fixed $\Psi_O$: circle (○) symbols

For the default scenarios there was seen a reduction in the atmospheric mass transport $Psi_a$ and therefore an instantaneous reduction in ocean mass transport $Psi_o$ through $\Psi_o/\Psi_a = 0.1$. With oceanic mass transport now independent of changes to the atmosphere the main difference shown is an increase in oceanic heat transport. Due to this strengthening of the ocean circulation there is much more heat being transferred from the Tropics to the Extra-Tropics, leading to a reduction in temperature gradient. Although the specific humidity gradient increases, this increase is less pronounced for a fixed $\Psi_o$ scenario, with values plateauing at $2 \times CO_2$ and higher concentrations. As a result, latent and atmospheric heat transports decreases, partially compensating for the relative increase in $H_o$. The magnitude change in $H_a$ is smaller than its oceanic counterpart thus leading to a relative increase in both the total heat
transport of the system and the ratio of $H_O/H_A$.

Ocean heat content (and therefore temperature) increases in 3 of the boxes ($T_{S2}, T_{O1}, T_{O2}$), with the tropical surface box $T_{S1}$ decreasing. These are attributed to the relative enhancement of ocean heat transport away from the tropics and into the rest of the ocean. The balancing of ocean temperature leads to a relatively unchanged global surface temperature.

Salinity follows a similar pattern to that of temperature, with increasing values in $S_{S2}, S_{O1}$ and $S_{O2}$, whilst $S_{S1}$ decreases in salinity. This is again due to the stronger ocean circulation advecting much more salt to higher latitudes, leading to both a relative and absolute decrease of the salinity gradient.

The ratio of the temperature and salinity gradients $\Gamma$ remains relatively unchanged apart from extremely small increases at the highest $CO_2$ concentrations.

There is an enhancement of net surface heat flux across both regions, indicating extra heat input in the Tropics and extra output in the Extra-Tropics due to increased surface temperatures. This slightly increases values for net TOA radiation meaning the system needs much more time to reach equilibrium. Surprisingly evaporative cooling radiation in box 1 decreases even though it is warmer; for box 2 there is a slight increase.
Figure C.4: Meridional heat transport for the ocean (cyan) and atmosphere (magenta) a), with total (ocean and atmosphere) system heat transport b). Atmosphere (red) and ocean (blue) circulation strength, with the ratio of the two equal to 0.1, consistent with model parameterization c). Rate of precipitation for the tropical (red) and extra-tropical (blue) region d). CO2 concentrations of fixed and transient evolutions are represented by symbols and curves respectively. Fixed variable scenarios are identified for fixed $\Psi_O$ (circles, dashed line), fixed $\Psi_A$ (triangles, dotted line), fixed $\Psi_O$ and $\Psi_A$ (squares, dashed-dotted line), and no fixed variables (stars, solid line). Curves in subplot d) have been smoothed using a 365 day filter to remove the stochastic forcing signal. Fixed CO2 symbols for subplot d) are the mean value for the last 1 year of the time series.

From the relation $F = \Psi_O/S_O$, freshwater transport is now directly proportional to the salinity gradient in equilibrium scenarios. F thus decrease with increasing $CO_2$ concentration.

The relatively small changes to DSE and L leave the ratio of the two terms $\gamma$ almost unchanged from default scenario values.

C.3.2 Fixed $\Psi_A$: triangle ($\Delta$) symbols

With a constant atmospheric mass transport, the decrease in $\Psi_a$ with $CO_2$ never happens. Relative to the default scenarios, much more heat is transported by the atmosphere from the tropical into the extra-tropical box. The increase in surface temperature for the region $T_{S2}$ is greater than for fixed $\Psi_O$ scenarios, leading to a larger decrease in the temperature gradient. Global mean temperatures rise more than the fixed $\Psi_O$ scenarios, by about 0.2°C at $2 \times CO_2$. The large changes in extra-tropical surface temperatures relative to that in the tropics, is high enough to decrease the specific humidity gradient to values below those found for the control run scenarios. Ocean heat content increases throughout the whole ocean as the high latitudes...
Figure C.5: Model atmospheric CO\textsubscript{2} concentration for fixed and transient scenarios a). Lower tropospheric specific humidity transport between each region b). Surface and atmospheric temperature difference between each region c). Atmospheric heat transport (magenta) decomposed into its Dry Static Energy (blue) and Latent (red) heat components d). CO\textsubscript{2} concentrations of fixed and transient evolutions are represented by symbols and curves respectively. Fixed variable scenarios are identified for fixed Ψ\textsubscript{O} (circles, dashed line), fixed Ψ\textsubscript{A} (triangles, dotted line), fixed Ψ\textsubscript{O} and Ψ\textsubscript{A} (squares, dashed-dotted line), and no fixed variables (stars, solid line). Curves in subplot b) have been smoothed using a 365 day filter to remove the stochastic forcing signal. Fixed CO\textsubscript{2} symbols for subplot b) are the mean value for the last 1 year of the time series.

Freshwater (and therefore latent heat) transport significantly decreases compared with the default and fixed Ψ\textsubscript{O} scenarios. For the previous two scenarios Ψ\textsubscript{A} increased whilst the specific humidity gradient decreased for increasing CO\textsubscript{2}, aiding in counteracting any significant shift in freshwater transport. With fixed Ψ\textsubscript{A} however, freshwater transport is directly coupled to the specific humidity gradient through \( L = \Psi_A \Delta q \), and so decreases with increasing CO\textsubscript{2}.

Although the atmospheric heat transport is coupled to the decreasing values of \( \Delta T \) an \( \Delta q \), it is the constant atmospheric mass transport that sustains a strengthened \( H_a \) relative to both default and fixed \( H_o \) scenarios (Fig. C.5d). The ratio of DSE to latent heat transport significantly increases with CO\textsubscript{2} as opposed to negatively changing for the default and Ψ\textsubscript{O} scenarios.

Ocean heat transport slightly decreases due to the flattening of the temperature gradient and the decrease in ocean mass transport. Atmospheric heat transport decreases the least out of the described fixed variable scenarios, with total heat transport increasing. For the reasons
Figure C.6: Meridional freshwater transport calculated directly in the atmosphere (red) and indirectly (green) from salt conservation arguments a). The ratio of ocean to atmospheric heat transport calculated directly (blue) and indirectly from salt conversation theory (red) b). The ratio of DSE to Latent heat transport in the atmosphere a.k.a. $\gamma$ c). The ratio of $\Delta T / \Delta S$ a.k.a. $\Gamma$ and the T/S slope d). $CO_2$ concentrations of fixed and transient evolutions are represented by symbols and curves respectively. Fixed variable scenarios are identified for fixed $\Psi_O$ (circles, dashed line), fixed $\Psi_A$ (triangles, dotted line), fixed $\Psi_O$ and $\Psi_A$ (squares, dashed-dotted line), and no fixed variables (stars, solid line).

just described, the ratio of ocean to atmospheric heat transport decreases.

Equilibrium values of tropical surface salinity decrease whilst extra-tropical surface and thermocline readings of salinity increase. The salinity gradient effectively decreases with increasing $CO_2$. These changes are in similar direction to that of fixed $\Psi_O$ scenarios but at a smaller fraction.

Net surface heat flux reduces the most out of all fixed $\Psi$ scenarios for both regions with less heat being absorbed (emitted) in box 1 (2). The net TOA radiation increases more than both the default and fixed $\Psi_O$ scenarios. Evaporative cooling has similar trends to the fixed $\Psi_O$ scenario (though not as large) with decreasing (increasing) evaporative cooling effects for box 1 (2).

C.3.3 Fixed Freshwater Transport $F$

For fixed values of freshwater transport $F$, many quantities are relatively unchanged relative to the default scenarios. The major differences involve latent heat transport and the salinity in the water. The change in atmospheric heat transport is insensitive to the fixed moisture
transport case; the effective increase in moisture transport therefore drives an increase in the ratio of DSE to latent heat transport.

There is seen a positive salinification of surface tropical waters with a corresponding freshening at higher latitudes. This reflects the sustained freshwater transport, enhancing the pattern of evaporation (precipitation) at low (high) latitudes. Consequently the salinity gradient increases the most out of all scenarios so far. By looking at $F = M\Delta S/S_o$, since moisture transport is now a constant, the decrease in oceanic mass transport must be perfectly balanced by a corresponding increase in $\Delta S$.

C.3.4 Fixed $\Psi_O$ and $\Psi_A$: square (□) symbols

Within this forced climatic scenario, many of the changes seen are the largest compared to any of the other aforementioned fixed variable scenarios.

For both fixed $\Psi_O$ and $\Psi_A$ there is no decrease in the mass transport of either the ocean or atmosphere with increased $CO_2$. This extreme scenario results in the largest increase in extra-tropical surface temperatures by approximately $2K$, with both increases in the total thermocline layer. Surface temperatures at the Tropics decrease slightly due to the increased advection of heat away from the region. This decreases the temperature gradient but also increases global mean surface temperatures to similar values (albeit slightly higher) from fixed $\Psi_A$ scenarios. Ocean heat content increases for the total ocean, with the largest changes seen at high latitudes.

Ocean heat transport is enhanced, however slightly smaller to scenarios with fixed $O$. This weakness is due to constant $\Psi_O$ decreasing the temperature gradient that $H_O$ is proportional to. For similar reasons, atmospheric heat transport increases relative to the default, but the change is smaller than that for fixed $\Psi_A$ scenarios. This is primarily due to the surface temperature decreasing the most out of all scenarios. Since $\Psi$ is fixed for both the ocean and atmosphere, it is this important gradient that is responsible for varying heat transport. Given these differences in both $H_o$ and $H_a$, the reduction in total heat transport is the smallest out of all scenarios.

The specific humidity gradient decreases the most, in parallel with the largest reductions of the temperature gradient. This in turn drives the largest reductions in latent heat transport even though mass transport is fixed. This increase in $H_A$ and decrease in moisture transport results in the largest increase in $\gamma = DSE/L$.

From the combination of both fixed $\Psi_O$ and $\Psi_A$, and a warmer climate, tropical surface
salinity values decrease the most with high latitude surface waters and the deeper thermocline increasing in salinity. This aids in the largest reduction in the salinity gradient which is now proportional to the decreased latent heat transport in the atmosphere. The ratio of temperature and salinity gradients decreases, being predominantly controlled by the larger reduction in the former rather than in the latter.

The net TOA radiation has the largest increase out of all fixed $\Psi$ scenarios. Net surface heat flux and evaporative cooling radiation both decrease (increase) for box 1 (2), though not as much as for fixed $\Psi_A$ scenarios.

**C.3.5 Summary of Fixed $\Psi$ and $CO_2$ Scenarios**

We have shown that the climate system can change dramatically (and in some cases, reverse trends normally shown with $CO_2$ forcing) by holding certain parameters constant. What is obvious is that the climate is multi-dependent on many variables with not one ‘master’ variable controlling the whole system. A summary of results for how all variables change with increasing $CO_2$ concentration for all fixed $\Psi$ scenarios is given in Table. C.2.

One of most robust features of the model across all fixed $\Psi$ scenarios is the increase in surface temperatures across both regions, leading to similar increases in thermocline temperature and global mean surface temperatures. The latter has a strong dependence on atmospheric mass transport, with the oceanic component playing a secondary role. As a result scenarios that have the smallest temperature gradients result in greater temperature homogenisation between low and high latitudes. The temperature gradient sets the atmospheric circulation strength and by architectural default, the ocean’s too ($\Psi_o/\Psi_a = 0.1$), driving the advection processes of moving temperature, salinity, moisture and heat to higher latitudes.

As global surface temperatures rise, so does the temperature of the ocean with much heat being stored within the deep thermocline layer. Heat storage thus increases for all ocean layers with ocean temperature not reaching equilibrium after 300 years of integration. This has obvious consequences to net surface heat flux, causing slight offsets in absorbed (emitted) radiation in the Tropics (Extra-Tropics) that is mirrored in a global imbalance of net TOA radiation.

The interplay between $\Psi_a$ and $\Delta T$ is non-linear; although atmospheric mass transport will increase with increasing temperature gradient, there will come a point where the increase in $\Psi_a$ will start to decrease the temperature gradient as it advects heat from the tropics. This stabilisation effect helps limit the range of polar amplification and of global mean surface
temperatures.

Salinity evolution across all scenarios reaches equilibrium with its surroundings within 300 years. Apart from the default (no fixed \( \Psi \)), the salinity gradient decreases for all fixed \( \Psi \) scenarios as the relative increase in \( \Psi_O \) advects the higher salinity waters from the tropics towards higher latitudes, overcoming the effects of freshening from increased precipitation. The latter effect is however greater than the advection of salty water into the high latitude region for the default setting, leading to the observed increase in \( \Delta S \).

Apart from scenarios where \( \Psi_O \) or \( \Psi_A \) is fixed, both the overturning circulation strength for the ocean and atmosphere decrease, again predominantly due to the reduction in the equator to pole temperature gradient. This has a leading effect on both decreasing ocean and atmospheric heat transport with decreases in the latter (former) being greatest with fixed \( \Psi_O \) (fixed \( \Psi_A \)). Total heat transport unsurprisingly decreases, though the largest decrease is seen with the default scenario, consistent with no artificial support for the reduction of atmospheric/ocean circulation strength.

Although the specific humidity gradient is somewhat proportional to the temperature gradient, its relationship is highly non-linear, resulting in a large spread in \( \Delta q \) across all fixed scenarios. For default and fixed \( \Psi_O \) scenarios, \( \Delta q \) increases with \( CO_2 \), whilst for fixed \( \Psi_A \) and both \( \Psi_O \) and \( \Psi_A \) constant, \( \Delta q \) decreases.

DSE transport in the atmosphere decreases for all scenarios, following its \( DSE \propto \Delta T^2 \) dependence. Since changes in the temperature gradient are larger than those for specific humidity, Latent heat transport decreases albeit it less than DSE due to its \( L \propto \Delta T \Delta q \).

The ratio of DSE/L exhibits both increasing and decreasing trends, the latter associated with fixed ocean, and ocean and atmospheric circulation, whilst the latter is attributed to fixed ocean-only circulation and the default scenario settings. Due to latent heat transport being dependent on \( \Delta q \), the different sign changes to the ratio are thus controlled by changes in the specific humidity gradient.

The ratio of the temperature and salinity gradient \( \Gamma \) has no unifying trend across all scenarios with some increasing (fixed \( \Psi_O \) and fixed \( \Psi_O \)) whilst others decrease (default and fixed \( \Psi_A \)). Apart from the default scenario whereby \( \Delta T \) (\( \Delta S \)) decreases (increases), both the salinity and temperature gradients decrease with increasing \( CO_2 \). The different amounts by how much each variable decrease varies across each fixed \( \Psi \) scenario, leading to the different trends observed.

A similar pattern to alternating changes in the ratio of ocean to atmospheric heat transport
Appendix C

\( H_O/H_A \) since this quantity is proportional to \( \Gamma \).

C.4 The effects of varying \( \alpha, \beta \) and \( \gamma \) under a 2 \( \times \) CO\(_2\) environment

For a doubling of CO\(_2\) there is a positive shift in global mean surface temperatures for all values of \( K_A \) due to the warmer climate. The temperature gradient, atmospheric mass transport and latent heat transport all decrease relative to the control CO\(_2\) environment, however the changes between the different CO\(_2\) environments becomes smaller with increasing values of \( K_A \). In other words, the coupling between CO\(_2\) concentration and each variable becomes progressively weaker as global mean surface temperatures reach 20°C or more.

The specific humidity gradient increases with a doubling of CO\(_2\) for all multiples of \( K_A \) although the surface temperature gradient decreases; again this highlights the non-linear relationship between temperature and specific humidity. Just as with \( \Delta q \), latent heat transport decreases for a doubling of CO\(_2\). This result is surprising since most model climate models predict an enhancement of the hydrological cycle and therefore an increase in latent heat transport. Within this model and under the solo varying parameter of \( K_A \), changes in latent heat transport with CO\(_2\) are more tightly coupled to atmospheric mass transport rather than the specific humidity gradient.

Both ocean, atmospheric and total heat transports decrease for a doubling of CO\(_2\) for all multiples of \( K_A \). In this new CO\(_2\) environment the ocean does not exhibit any trend reversals seen previously, remaining positive throughout. The ratio of \( H_O/H_A \) still preserves a similar distribution, though now with a maximum value at the 0.2 multiple of \( K_A \).

For a doubling of CO\(_2\) all the quantities mentioned above increase in global mean surface temperatures relative to the default CO\(_2\) scenario. Unlike the perturbed \( K_A \) scenarios, the distribution of points for these new scenarios are shifted to the right (increasing \( T_s \)) and vertically (increasing/decreasing values) but to such an extent that they overlap the linear ‘space’ of the control CO\(_2\) scenario points.

The largest differences between default and 2 \( \times \) CO\(_2\) scenarios occur at \( \alpha \) multiple of 0.5 where global mean surface temperature change is approximately 3K. Just like in the perturbed \( K_A \) scenarios, the effects of CO\(_2\) concentration for the larger multiples of 5.0 and 10.0 are weak, with both scenarios having extremely similar values with small changes in \( T_s \).

For a doubling of CO\(_2\) all quantities exhibit a positive shift in global mean surface temper-
atures for all multiple ranges of $\gamma$, consistent with the warmer climate. Values of the specific humidity gradient increases with $CO_2$ across all multiple ranges, whereas the temperature gradient (and therefore atmospheric mass transport) decreases. A surprising feature for these perturbation scenarios is that latent heat now increases with $CO_2$ concentration for multiples of $\gamma < 1.0$, where above this multiple, latent heat transport then decreases. From the increased resolution in $\gamma$ in Fig. ??, this critical multiple value happens at $0.6 \times \gamma$, with multiples less than (greater than) than this value constituting a climate where latent heat transport increases (decreases) with increasing $CO_2$ concentration. At $0.6 \times \gamma$, $T_s \approx 7^\circ C$. This model attribute should not be confused with hysteresis whereby a model can exhibit more than one distinct equilibrium for the same starting conditions; here we are varying the parameter $\gamma$ each time thus differing our starting conditions for every integration period.

For multiples of $\gamma$ equalling 0.1 and 0.2, ocean, atmosphere and total heat transports increase with a doubling of $CO_2$, whilst multiples above the latter value decrease. Though both $H_o$ and $H_a$ are proportional to $\Delta T_s$, the larger decrease in ocean heat transport as compared to its atmospheric counterpart at $0.5 \times \gamma$ is partially due to the lower mass transport of the ocean. The ratio of ocean to atmospheric heat transport also decreases with a doubling of $CO_2$, though the differences becoming progressively smaller at high multiples of $\gamma$.

Although we have found a regime in the model where latent heat transport does increase with $CO_2$ concentration, this effect is only seen in cold climates i.e $T_s < 7^\circ C$. Many of the latest GCM models predict an increase of latent heat transport at mid-latitudes as the climate warms, however these are in the $T_s$ range of $13^\circ C$ or larger. The EPcm model therefore does not mirror the same surface temperature range for latent heat enhancement to take place.
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Table C.2: The changes to variables under increased \( CO_2 \) for different fixed \( \Psi \) scenarios. Up (down) arrows represent increasing (decreasing) trends. Asterix signs represent under which fixed scenario shows the greatest changes. \textit{Heat Flux} is abbreviated as \( h.f. \). Dashed (-) symbols represent no change to variable due to it being already fixed within the scenario i.e. for \( \Psi \).
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