# Abstract

## Declaration

I confirm that this is my own work, and the use of all material from other sources has been properly and fully acknowledged.

David Sproson.


## Acknowledgements

I would like to thank my supervisor, David Marshall, for all of his help and support over the last few months, for many interesting discussions and for suggesting an interesting topic in the first place.

## Symbols, Abbreviations and Acronyms

$\alpha$ - Thermal expansion coe□ecient

$\beta$ - Saline expansion coe□cient or meridional derivative of the Coriolis frequency

□ - Energy dissipation per unit mass

$\Gamma$ - Mixing e□ciency

$\kappa$ - Di□usivity

$\mu$ - Dynamic coe□cient of viscosity

$\nu$ - Molecular coe□cient of viscosity

$\rho$ - Density

$\tau$ - Surface wind stress. In vector form this is the 3-D wind stress vector consisting of components $\tau$ , $\tau_y$, $\tau_z$. In scalar form it is the zonal wind stress

A - f437205(s)-0.649399(h)[illegible]

hciest[illegible]s

**p** - **Pressure**

**q** - **Volume Transport**

$\mathbf{Q}_n$ - **Northern hemisphere sinking**

$\mathbf{Q}_s$ - **Southern Ocean upwelling**

**Q** - **Low latitude upwelling**

**S** - **Salinity in parts per thousand**

**T** - **Temperature**

**THC** - **Thermohaline circulation**

**u** - **3-D velocity vector consisting of components u, v, w**

# Contents

1 Introduction 4
1.1 The Thermohaline Circulation . . . . . . . . . . . . . . . . . . . . . . 4
1.2 Sandström's Theorem . . . . . . . . . . . . . . . . . . . . . . . . . . . 5
1.3 Outline of this Study . . . . . . . . . . . . . . . . . . . . . . . . . . . 6

2 Energetics of the General Circulation 8
2.1 The Requirement for Mixing Energy . . . . . . . . . . . . . . . . . 8

3.1.4 Volume Balance . . . . . . . . . . . . . . . . . . . . . . . . . 33
3.1.5 Model Parameters . . . . . . . . . . . . . . . . . . . . . . . . 33
3.2 Solutions in Limiting Scenarios . . . . . . . . . . . . . . . . . . . . . 35
3.2.1 In the Limit of no Southern Ocean Processes . . . . . . . . . 35

# List of Figures


1.1 The Thermohaline Circulation . . . . . . . . . . . . . . . . . . . . . . 5

2.1 Potential density section in the Pacific Ocean . . . . . . . . . . . . 11
2.2 Comparing boundary and uniform mixing in a GCM . . . . . . . . 16
2.3 Wind input into the general circulation . . . . . . . . . . . . . . . . 22

3.1 Schematic of Gnanadesikan's model . . . . . . . . . . . . . . . . . . 31
3.2 Pycnocline depths when K is neglected . . . . . . . . . . . . . . . 36
3.3 Pycnocline depths when $Q_n$ is neglected . . . . . . . . . . . . . . . 38
3.4 τ against K for the Full Gnanadesikan Model . . . . . . . . . . . 40
3.5 A against K for the Full Gnanadesikan Model . . . . . . . . . . . 41
3.6 Pycnocline depth varying with a stochastic wind stress . . . . . . 42
3.7 A plot of Gnanadesikan's model function . . . . . . . . . . . . . . . 43

4.1 Two box model of the t.h.c. (Stömmel, 1961) . . . . . . . . . . . . 46
4.2 Equilibria in a two box model of the t.h.c. . . . . . . . . . . . . . . 48
4.3 Results of the reparameterized two layer model 1 . . . . . . . . . . 53
4.4 Results of the reparameterized two layer model 2 . . . . . . . . . . 54
4.5 Results of the reparameterized two layer model 3 . . . . . . . . . . 55
4.6 Results of the reparameterized two layer model 4 . . . . . . . . . . 56
4.7 A time dependent experiment showing multiple equilibria . . . . . 58

# Chapter 1

# Introduction

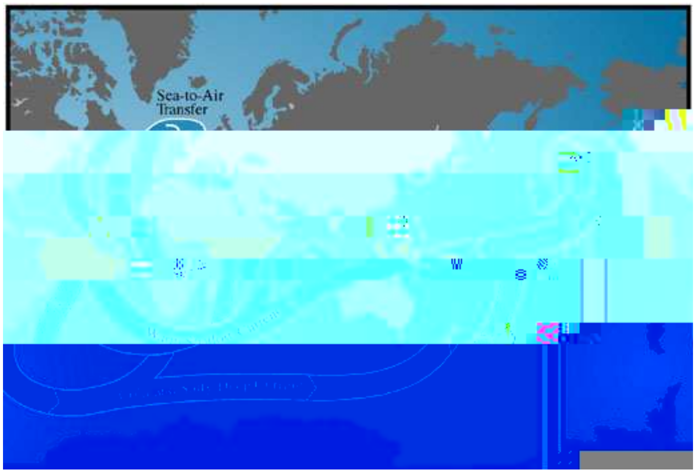



**Figure 1.1: A highly simplified schematic of the thermohaline circulation. This conveyor belt visualization of the thermohaline circulation was put forward by**

which processes are dominant in supplying energy to the oceans.

Having identified the processes most important for providing energy to the thermohaline circulation, in Chapter 3 some original calculations are presented using a simple model of the thermohaline circulation which incorporates these processes. The nature of the circulation and depth of the pycnocline are investigated under various parameter regimes. Chapter 4 examines multiple equilibria in the thermohaline circulation and addresses the paradox of multiple equilibria in a mechanically driven ocean when traditional ideas of multiple equilibria rely on opposing thermal and haline forcing.

,n,Fi Finally in Chapter 5, a summary of the study is given, the 26269(i)2(e)3.56148(r)-2.26269(.)2

# Chapter 2

# Energetics of the General Circulation

## 2.1 The Requirement for Mixing Energy

dissipation: If, in an experiment on turbulent flow, all of the control parameters are kept the same, except for viscosity, $\nu$, which is lowered as much as possible,

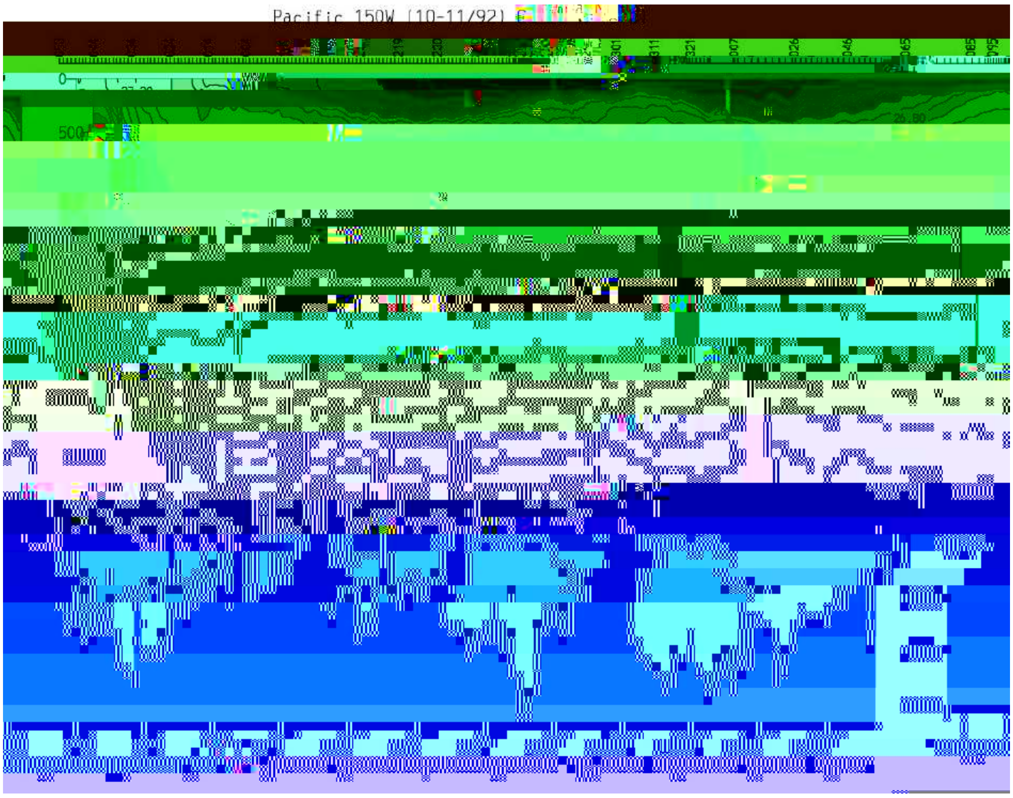


Figure 2.1: Meridional neutral density section in the central Pacific Ocean, from Wunsch & Ferrari (2004). Notice that below the pycnocline (at approximately 700 m) and away from the Southern Ocean upwelling the isopycnals are approximately horizontal. This led Munk (1966) to introduce a one-dimensional balance between advection and diffusion as a process for maintaining the stratification.

sea floor, and rewriting this as

$$\int_A \mathbf{u}\theta \cdot \hat{\mathbf{n}}\, d$$

un-physically, confined to $x < 0$. The solution is omitted here (however see Munk & Wunsch (1998), Appendix B). Where $u = 0$, (2.9) is reduced to the Munk (1966) balance between vertical advection and di□usion, which has a scale depth of approximately 100 m associated with di□usivity $\kappa_{\mathrm{PE}}$ over the entire ocean basin. As $u \to \infty$, the basin average di□usivity tends to the boundary value of $\kappa = 10^{-4}$ $m^2s^{-1}$ giving a scale depth of 1 km. Assuming a typical basin width of 10000 km, it was shown that the solution approached the $u = 0$ case for $u \leq O(0.1)$ $mms^{-1}$ and the $u \to \infty$ case for $u \geq O(1)$ $mms^{-1}$. Typical mesoscale circulation rates in the ocean are usually well above the 1 $mms^{-1}$ required for the solution to approximate the $u \to \infty$ case. The upshot of this is that it seems that horizontal advection is capable of transporting T,S characteristics into the ocean interior. Of course, this model is far from realistic, not only in its lack of dimensionality but also in its treatment of u as a constant everywhere in the domain, which is clearly not the case in reality. However, despite these shortcomings, the model may provide an adequate description of the process by which lateral homogenisation occurs.

Marotzke (1997) considered the e□ects of boundary processes on a full ocean general circulation model (GCM) forced simply by a tendency to restore temperature and ocwn

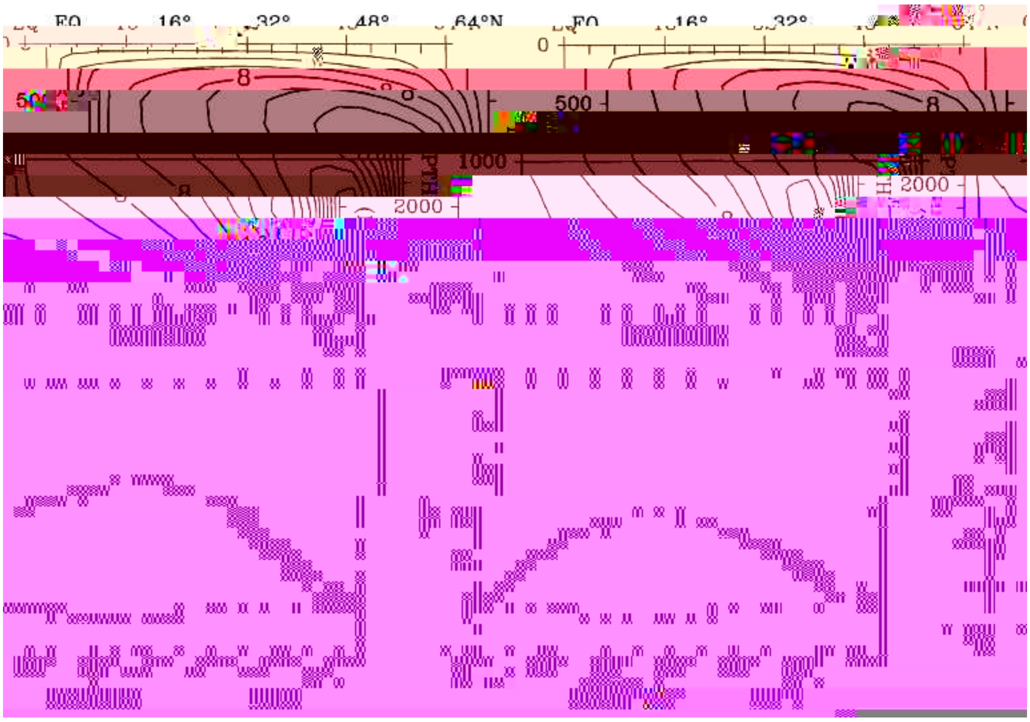



**Figure 2.2: Comparing the meridional overturning circulation (a) and meridional heat transport (b) for boundary mixing (left) and unifo4.6091(E)3.71873(N)1.2b-376.549(m)2.962t**

# CHAPTER 2. ENERGETICS OF THE GENERAL CIRCULATION

The former will be denoted $\Phi_{\mathrm{tide}}$. The internal energy balance for the same fluid parcel is

$$\frac{\partial}{\partial t}(\rho I) + \nabla \cdot \left[ \rho I \mathbf{u} + \mathbf{F}_{\mathrm{rad}} - \rho c_p \kappa \nabla T - \frac{\partial h_E}{\partial S} \rho \kappa \nabla S \right] = -p \nabla \cdot \mathbf{u} + \rho \epsilon, \quad (2.11)$$

where $I$ is the internal energy and $\mathbf{F}_{\mathrm{rad}}$ represents any radiative flux between the atmosphere and ocean, where a flux from the atmosphere to the ocean is positive, which occurs near the surface. The $\rho c_p \kappa \nabla T$ term simply represents the irreversible diffusion of heat, with $c_p$ the specific heat of sea water at constant pressure. The final gradient term on the left hand side, $(\partial h_E/\partial S)\rho\kappa\nabla S$, is the salinity diffusion term. This represents the generation of heat in the fluid due to solution energy of salt, with $h_E = I + \rho/p$. Note that there are two terms which occur on the right hand side of both (2.10) and (2.11) with opposite sign. These are the routes of conversion between kinetic and internal energy. The $p\nabla \cdot \mathbf{u}$ term, known as the 'compressive work' is the main, reversible route of conversion between kinetic and potential energy, while $\rho\epsilon$ represents the generation of heat through the viscous dissipation of kinetic energy (i.e. friction).

The potential energy balance is given by

$$\frac{\partial}{\partial t}(\rho\Phi) + \nabla \cdot (\rho\Phi\mathbf{u}) = \rho\mathbf{u} \cdot \nabla\Phi + \rho\frac{\partial \Phi_{\mathrm{tide}}}{\partial t}, \quad (2.12)$$

where $\Phi$ is as before. The $\rho\mathbf{u} \cdot \nabla\Phi$ term is also present in (2.10), but with an opposite sign. This represents a reversible path between kinetic and potential energy – the route which must be taken for turbulent mixing to maintain the oceanic stratification and permit the existance of the thermohaline circulation.

The equations (2.10) – (2.12) can be converted to global energy budgets by integrating over the global ocean and applying Stokes' Theorem. This gives

$$\begin{aligned} \frac{\partial}{\partial t} \iiint \rho E \, dV = &- \iint_A \rho E \, (\mathbf{u} - \mathbf{u}_s) \cdot \hat{\mathbf{n}} \, dA - \iint_A (p\mathbf{u} + \mu\nabla E) \cdot \hat{\mathbf{n}} \, dA \\ &- P + C - D, \end{aligned} \quad (2.13)$$

$$
\begin{aligned}
\frac{\partial}{\partial t}\iiint \rho I \, dV = &-\iint_A \rho I\,(\mathbf{u}-\mathbf{u}_s)\cdot\hat{\mathbf{n}}\,dA \\
&-\iint_A \left(\mathbf{F}_{\mathrm{rad}} - \rho c_p \kappa_T \nabla T - \frac{\partial h_E}{\partial S}\rho\kappa_S \nabla S\right)\cdot\hat{\mathbf{n}}\,dA - \mathcal{C} + \mathcal{D},
\end{aligned}
\tag{2.14}
$$

$$
\frac{\partial}{\partial t}\iiint \rho\Phi \, dV = -\iint_A \rho\Phi\,(\mathbf{u}-\mathbf{u}_s)\cdot\hat{\mathbf{n}}\,dA + \iiint \rho\frac{\partial \Phi_{\mathrm{tide}}}{\partial t}\,dV + \mathcal{P} \tag{2.15}
$$

where $\mathbf{u}_s$ is the free surface velocity. The script characters $\mathcal{P}$, $\mathcal{C}$ and $\mathcal{D}$ are the volume integrals of the conversion terms in (2.10)–(2.12) and so are energy transformations rather than sources,

$$
\mathcal{P} = \iiint \rho\mathbf{u}\cdot\nabla\Phi \, dV, \quad \mathcal{C} = \iiint p\nabla\cdot\mathbf{u}\, dV, \quad \mathcal{D} = \iiint \rho\epsilon \, dV.
$$

The first term on the right hand side of each of the equations (2.13)–(2.15) represents advection through the free surface. Making the fairly safe assumption that the ocean is in an approximate steady state, we must have that $\iint_A (\mathbf{u}-\mathbf{u}_s) \approx 0$. Therefore, for any energy to be advected across the surface there must be a correlation between $E$, $\Phi$

momentum forcing in (2.13), tidal forcing in (2.14) may be lateral side boundaries can be written as

$$\iint_A dA \left[\overline{\rho E w + p w}\right]_{z_1}^{z_2} = -\iiint dV\, \overline{\rho g w - p \cdot}$$

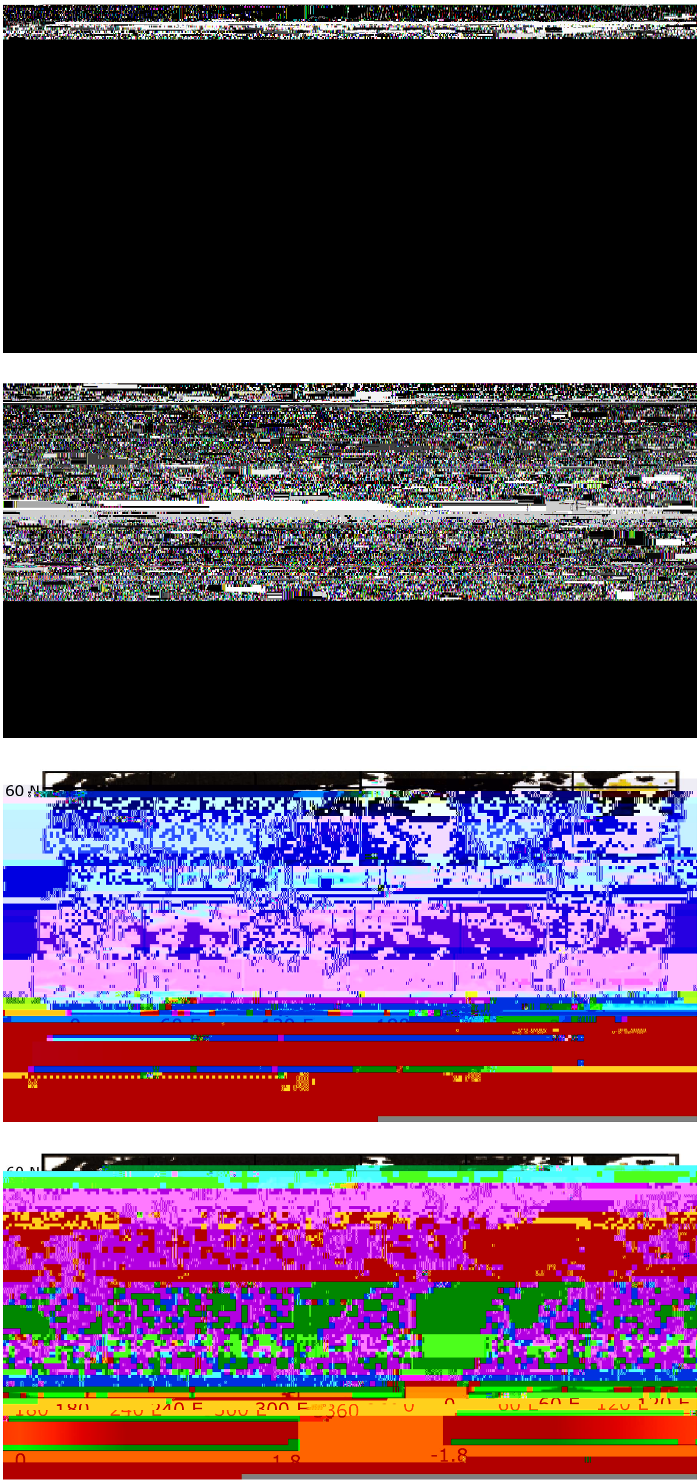


**Figure 2.3: Four year average wind input and deviations from it. (a) $\langle \boldsymbol{\tau} \rangle \cdot \langle \mathbf{u}_g \rangle$, (b) $\langle \boldsymbol{\tau}_y \rangle \cdot \langle \mathbf{v}_g \rangle$, (c) $\langle \boldsymbol{\tau}' \mathbf{u}'_g \rangle$, (d) $\langle \boldsymbol{\tau}'_y \mathbf{v}'_g \rangle$. All units are $\mathrm{Wm}^{-2}$. Adapted from Wunsch (1998).**

the time averaged scalar product of surface horizontal wind stress, $\boldsymbol{\tau}$, and surface horizontal geostrophic velocity, $\mathbf{v}_g$. Due largely to a lack of data, many of these studies have been very crude, and hence are likely to be inaccurate. For example, the study in 1994 by Oort et al relied on ship drift data to compile a geostrophic velocity and wind climatology to obtain wind stresses. This study, as that of Fofonoff before, arrived at a value 2 TW of work on the general circulation (a huge amount when one considers that only just over 2 TW of mechanical mixing energy is required to sustain the circulation).

Wunsch (1998) used the same approach, but split the work into mean and fluctuating parts,

$$W = \langle\boldsymbol{\tau}\rangle \cdot \langle\mathbf{v}_g\rangle + \left\langle \boldsymbol{\tau}' \cdot \mathbf{v}_g' \right\rangle, \tag{2.23}$$

How does the wind impart energy to the general circulation? Wunsch & Ferrari (2004) noted that one cannot define a unique pathway by which atmospheric kinetic energy is transferred into the oceans. Any wind stress on the surface will cause in an Ekman process, generating a finite vertical velocity, w$_E$. This results in an upward or downward deflected isopycnal which will raise or lower the potential energy of the ocean respectively. However, by geostrop399(i)0.972056(5671(65)3.56140

# CHAPTER 2. ENERGETICS OF THE GENERAL CIRCULATION

# CHAPTER 2. ENERGETICS OF THE GENERAL CIRCULATION

In the time average, water evaporates in the tropics, lowering the centre of mass of the ocean here, with a corresponding amount of precipitation increasing the potential energy in the high latitudes through an elevation in the free surface. During its transport, however, the fresh water loses heat to the atmosphere, the overall effect of this process is to lower the energy of the oceans. Huang (1998) estimated the net work through fresh water transport at $-10^{-2}$ TW as a global average. Again, due to the large relative uncertainties involved, this is generally approximated as zero when making energy budgets.

### 2.4.5 Atmospheric Pressure Loading

The free surface of the ocean responds to changes in atmospheric pressure loading.

required for the maintenance of the stratification is provided from drag of the the general circulation, 0.1 TW directly and 0.1 TW from mesoscale eddies. It should be noted that the figures we are using here are just the generally accepted values, and the uncertainties are hopelessly large – “at d056(e)-265409

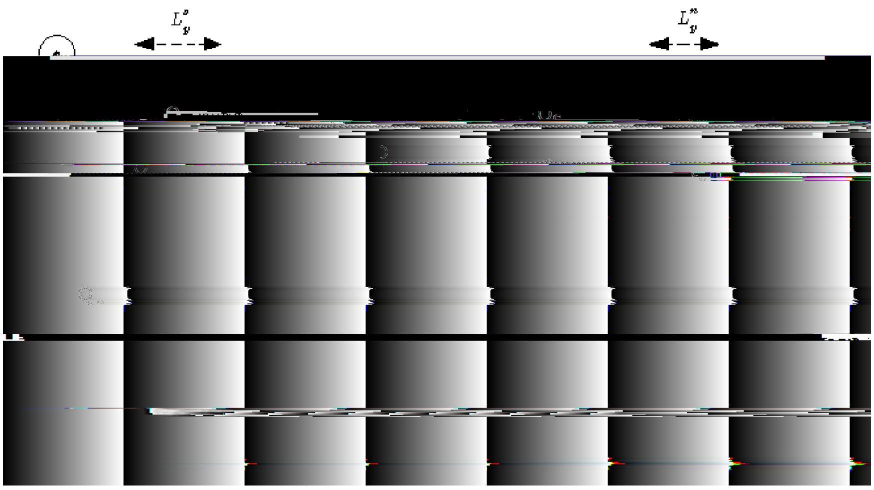
# Chapter 3

# A Two Layer Model of the Thermohaline Circulation

In the previous sections we have seen that two of the most important factors in controlling the thermohaline circulation are wind input, most of which goes into the Southern Ocean, and vertical mixing which is essential to maintain the oceanic stratification. We now present some original calculations and analysis of a variation of a model first proposed by Gnanadesikan (1999), which includes both of these factors in a simple and highly idealized model of the thermohaline circulation, in an attempt to gain a better understanding of how there factors affect the circulation. The model involves three key processes, namely northern hemispheric sinking, low latitude upwelling and an Ekman driven upwelling in the Southern Ocean, partially balanced by an eddy return flow. These processes are parameterized in terms of the depth of the pycnocline, an area of the ocean where density rapidly increases with depth, and which generally delineates the surface mixed layer from the abyssal ocean.

## 3.1 Formulation

The ocean is assumed to consist of two layers, a surface layer and the abyssal ocean which are separated by the pycnocline. The setup Gnanadesikan used, shown in Figure 3.1, was justified by an observational section of the potential density in the Atlantic Ocean. Throughout most of the ocean, the pycnocline is approximately horizontal (constant depth), extending to the surface in an ap-


$L_y^s$
$L_y^n$

km, and A = $2.44 \times 10^{14}$ $m^2$. The Southern Ocean wind stress, $\tau$, is of course variable, however it is believed to lie between 0.1 and 0.2 $Nm^{-2}$. In this study values of $\tau$ ranging from 0 to 0.3 $Nm^{-2}$

## 3.2 Solutions in Limiting Scenarios

**In a steady state, the model, as shown in (3.9), is a cubic poly**

and we can divide through by $D$ to obtain the quadratic eqaution

$$\frac{g'}{2f}D^2 + \frac{A\ L}{L_y^s}D - \frac{\tau L}{\rho f} = 0. \tag{3.12}$$

Note that by doing this we lose one of the solutions to (3.11), however this is simply the solution where $D = 0$, which in the case of $K = 0$ implies no overturning circulation. As (3.12) is quadradic, we can find two solutions, $D_1$ and $D_2$, where

$$D_{1\,2} = \frac{-A\ L_x}{L_y} \pm \sqrt{\left(\frac{A\ L_x}{L_y}\right)^2 + \frac{2g'\ L_x}{f^2}}$$

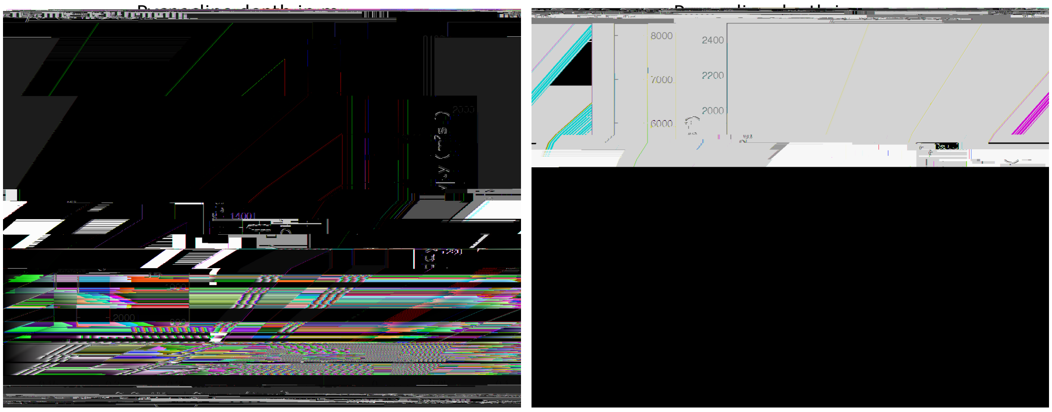


Figure 3.3: Pycnocline depth in the

in the Southern Ocean and low latitudes is by a significant deepening of the pycnocline. This deepening also reduces the upwelling in the low latitudes. Figure 3.3 shows the depth of the pycnocline when $Q_n \equiv 0$ in the $\tau$, $A$ parameter space. Notice that in much of the space the pycnocline obtains an unrealistically large depth. This is necessitated in order to remove fluid from the surface layer in the limit $\tau/\rho f \rightarrow A/L_y^s$. Figure 3.3 also shows that when $K$ is small, the pycnocline depth is more sensitive to wind stress, particularly when this is also small.

# CHAPTER 3. A TWO LAYER MODEL OF THE THERMOHALINE CIRCULATION

## 3.6 Summary

Gnanadesikan (1999) proposed a simple two layer model of the thermohaline circulation. By identifying and parameterizing a small number of processes key to the existence of the circulation in terms of the depth of the pycnocline, $D$, fluid transports could be found by solving a cubic equation in $D$.

The solutions and behaviour of this model were investigated analytically in limiting scenarios and numerically for the full equation. It was shown that when Southern Ocean processes are neglected the circulation strength follows a power law found in other work involving numerical general circulation models, for example Marotzke (1997), Bryan (1987). It is unclear whether or not this is coincidental. When interior upwelling was neglected, it was found that, in the limit of no eddy return flow, the circulation was controlled by the Ek.836(f)-152(l)(y)-396.632(t)34(o)-3

# Chapter 4

# Multiple Equilibria in the Thermohaline Circulation

We have seen in the previous chapter that the model of the thermohaline circulation proposed by Gnanadesikan (1999) has only one equilibria for a realistic

of these boxes is given by

$$|q|S_1 = (|q| + E)\,S_2,$$

or equivalently

$$|q|\Delta S = ES_2 \approx ES_0.$$

The modulus signs are needed here as no assumptions have been made about the direction of the circulation. Assuming that the transport is a linear function of the density difference, we have

$$q = k\,(\alpha\Delta T - \beta\Delta S),$$

and thus

$$|q|q - k\alpha\Delta T|q| + k\beta ES_0 \approx 0. \tag{4.1}$$

The strength of the circulation is thus given by a quadratic equation in q, the transport. In the case where $q > 0$, (4.1) becomes

$$q^2 - k\alpha\Delta Tq + k\beta ES_0 \approx 0,$$

which has two real root for

$$\frac{k\alpha^2\,(\Delta T)^2}{4\beta S_0} > E.$$

Assuming standard values of $\alpha = 2 \times 10^{-4}$ $K^{-1}$, $\beta = 0.8 \times 10^{-3}$ $({}^0/_{00})^{-1}$, $k = 0.5 \times 10^{10}$ $m^3s^{-1}$, $\Delta T = 20$ K, $S_o = 35$ ${}^0/_{00}$, two real solutions exist when

$$E < 0.7 \text{ Sv (approx.)},$$

and are given by

$$\frac{k\alpha\Delta T \pm \sqrt{k^2\alpha^2\,(\Delta T)^2 - 4k\beta ES_0}}{2}. \tag{4.2}$$

When $E = 0$, (4.2) reduces to

$$\begin{aligned} q &= 0,\ k\alpha\Delta T \text{ Sv} \\ &= 0,\ 20 \text{ Sv}. \end{aligned}$$

In the case where $q < 0$, there exists only one real solution to (4.1). The


**Figure 4.2: Solution of (4.1) showing multiple stable equilibria and a mechanism by which the thermohaline circulation could collapse, jumping from one stable equilibrium to the second. Solid lines show the stable equilibria and the dashed line represents an unstable equilibria.**

**combined solutions are shown in Figure 4.2. Here the solid lines represent stable msematemi modes of the thermohaline circulation and the dotted line represents an unstable mode. The upper solid line represents the state the thermohaline circulation is in currently, whereby a large amount fluid sinks in the northern hemisphere driven by temperathgPT**

sity gradients in the fluid. Such a system, as argued by Sandström and others, can only remove energy from the fluid, and thus cannot maintain a circulation. However, when one looks back at the assumptions which have been made, the differences between the two box, the two layer and the energetic ideas are not so vast. One of the key assumptions made for the two box model is that both of the boxes are well mixed. Such an assumption means that a turbulent field within the ocean is being implied, but no energy source is related to provide this. From the earlier discussions we know that the energy required for this is provided primarily by the tides and the winds, which clearly cannot be included in such a simple model. This situation is quite similar to that in the two layer model considered earlier. Then a constant upwelling in the low latitudes, $K$ , was assumed, again implying some turbulent field, and again without any associated energy source.

## 4.2 Multiple Equilibria in General Circulation Models

One of the first investigations into multiple equilibria in the general circulation using a coupled atmosphere-ocean general circulation model was carried out by Manabe & Stouffer (1988). They started the integration from a state where the atmosphere was dry, isothermal and at rest, and the the ocean was isothermal, at rest and had a uniform salinity. Once the model had reached a steady state, Manabe & Stouffer conducted two experiments. In the first they forced the model by restoring the surface salinities to the observed profiles on a timescale of 30 days, while in the second this forcing was not included. Two quasi-stable equilibria were found which were qualitatively similar to the two states found in the Stömmel (1961) model. The solution where the thermohaline circulation exists arose from the first experiment where surface salinities were restored, and the second equilibria, where there is no large overturning in the north Atlantic was simply achieved by integrating the model forward from the aforementioned initial conditions. That the existence of a thermohaline circulation requires the extra forcing comes as no great surprise. The sinking of the thermohaline circulation in the northern Atlantic is controlled largely by sinking of water that has become dense due to evaporation, which leaves the remaining wateroibnhhen halinity-2.26269(l)0.979493]TJ-

**between the low latitudes and the southern hemisphere. In the two box model, the collapse occurs when the northern sinking falls below ap**

becomes a sinking of around 4 Sv, balanced by an upwelling of similar magnitude in the low latitudes, which has reduced only slightly. In the collapsed state the pycnocline depth becomes very sensitive to the Southern Ocean eddy diffusivity. This happens because in the absence of northern sinking, the only process which can remove fluid from the surface layer is the Southern Ocean eddy return flow. As the eddy diffusivity is lowered, the pycnocline depth must increase to allow a similar amount of fluid to be returned to the abyssal layer.

Figure 4.4 shows the steady-state solutions in the $\tau$, K parameter space. Again the solutions are for the northern sinking solution where it exists, which is everywhere other than where $\tau$ and K are suitably small. Again in the collapsed state a sinking of between 5 Sv and 10 Sv in the southern hemisphere is balanced by a similar upwelling in the low latitudes. In the collapsed state the pycnocline depth is very sensitive to the wind stress and increases with it, a response necessary to allow the removal of the extra fluid upwelling into the surface layer.

Figures 4.5 and 4.6 show the steady state solutions in the A , K and $\tau$, K parameter spaces when the integration was started in the collapsed state ($D_0$ = 1 m). The fact that these figures differ from Figures 4.3 and 4.4 shows that the reparameterization of $Q_n$ has indeed introduced multiple equilibria into the model, rather than simply scaling out the northern sinking for certain sets of parameters and allowing it to recover if we then revert to the standard parameter set. This is better seen through the time dependent problem, which we will come to shortly. In the A , K parameter space, the solutions do not vary a great

# CHAPTER 4. MULTIPLE EQUILIBRIA IN THE THERMOHALINE CIRCULATION

**circulation, it did not support multiple equilibria which are observed in many other analytic and numerical ocean models.**

**An investigation into a version of the two box model of the thermohaline circulation proposed by Stömmel (1961) gave an approximate value that the fluid**

# Chapter 5

# Summary and Discussion

Sandström's Theorem, together with the no turbulence theorem of Paparella & Young strongly suggest that the ocean is not forced by buoyancy exchange at the surface, but rather that the energy input required to maintain the circulation comes from other sources in the from of mechanical mixing energy which maintains that stratification against the circulation. This, of course, dispels the notion of a thermohaline circulation driven by freshwater transport and heat exchange with the atmosphere.

A number of observational campaigns have been carried out to attempt to quantify how much mixing energy is required to maintain the stratification. A dichotomy arose here between those studies that implied diapycnal mixing rates by measuring flow into and out of basins and those that used microstructure measurements to measure diapycnal mixing directly. The former found the required diapycnal di□usivity to be of the order of 10$^{-4}$ m$^2$s$^{-1}$, while the latter found rates of only 10$^{-5}$ m$^2$s$^{-1}$ over most of the ocean, a problem which became known as the missing mixing problem. The current theories suggest th

Although a vast amount of work is done by the winds on the ocean totaling some 20 TW, most of this remains in the surface layer and does not penetrate into the abyss where it would be available for deep ocean mixing.

The largest problem with energetic studies of the oceans is the lack of appropriate observations. The launch of satellite missions such as TOPEX/POSEIDON have provided a much greater coverage of the ocean surface, however observations in the ocean interior are still incredibly sparse, coming mainly from field campaigns which can only ever cover a tiny proportion of the ocean. This lack of observations causes the huge uncertainties in the oceanic energy budgets. Although it is possible to fill some of the gaps with model output, it is difficult to rely on this due to the lack of verification. It is also difficult to trust model results when conducting studies of energetics. This is due to the way that the stratification is maintained in general circulation models, which is usually done through to parameterize diapycnal mixing any more accurately and so the introduction of potential energy into the circulation must be accepted. These processes were parameterized in terms of the depth of the pycnocline, resulting in a cubic equation. This model, though very simple, captures qualitatively, and to a certain degree behind multiple equilibria we turned to a two box model of the thermohaline circulation first proposed by Stömmel (1961). It was seen that salinity gradients caused by freshwater transport could cause a reversal in the circulation, which in this semi-hemispheric model caused a sinking in the low latitudes, which is probably an unphysical result due to the intense heating which occurs in tropical regions. This collapsed state existed for all non-negative values of freshwater transport, showing that once the circulation has collapsed there is no reason for it to revert to a northern sinking state if the climate is restored to its previous conditions.

Using the results from the Stömmel (1961) box model, an attempt was made to incorporate multiple equilibria into the Gnanadesikan (1999) two layer model. This was done by reparameterizing the northern sinking term into an implicit form whereby once the northern sinking weakened to a certain point it was effectively scaled down to zero, simulating the collapse of the circulation due a salt feedback. The threshold for this collapse was taken to be similar to that in the two box model at around 10 Sv. This scaling may lead one to believe that, unlike other models which exhibit multiple equilibria, the circulation would recover when whatever parameter was changed to cause the collapse is retu(C)4.53967(U)1.27838(S

# BIBLIOGRAPHY

[24] Sandström, J.W. 1908. Dynamicsche Versuche mit Meerwasser. Annalen der Hydrographie under Martimen Meteorologie 36:6–23

[25] Scott, J.R., Marotzke, J., Adcroft, A. 2001. Geothermal Heating and its Influence on the Meridional Overturning Circulation. J. Geophys. Res. 106:3141–3154

[26] Scott, J.R., Marotzke, J., Stone, P.H. 1999. Interhemispheric thermohaline circulation in a coupled box model. J. Phys. Oceanogr. 29:351–365

[27] St. Laurent, L., Garrett, C. 2002. The Role of Internal Tides in Mixing the Deep Ocean. J. Phys. Oceanogr. 32:2882–2899

[28] Stein, C.A., Stein, S. 1992. A Model for Global Variation in Oceanic Depth and Heat Flow with Lithospheric Age. Nature. 359:123–129

[29] Stömmel, H. 1961. Thermohaline Convection with Two Stable Regimes of Flow. Tellus. 13:131–126

[30] Toole, J.M., Polzin, K.L. Schmitt, R.W. 1994. Estimates of Diapycnal Mixing in the Abyssal Ocean. Science. 264:1120–1123

[31] Toggweiler, J.R., Samuels, B. 1998. On the Ocean's Large Scale Circulation Near the Limit of No Vertical Mixing. J. Phys. Oceanogr. 28:1832–1852

[32] Wunch, C., Ferrari, R. 2004. Vertical Mixing, Energy, and the General Circulation of the Oceans. Annu. Rev. Fluid Mech. 36:281–314

[33] Wunsch, C. 1998. What is the Thermohaline Circulation? Science. 298:1180–1181

[34] Wunsch, C., Stammer, D. 1997. Atmospheric Loading and the "Inverted Barometer" E□ect. Rev. Geophys. 35:79–107

[35] Wunsch. C. 1998. The Work Done by the Wind on the Oceanic General Circulation. J. Phys. Oceanogr. 28:2332–2340