The effect of an increased freshwater flux on the meridional overturning circulation of the North Atlantic: A numerical study

Master thesis in Geosciences Meteorology and Oceanography

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Abstract

As global warming causes both glaciers and sea ice to decay, increased freshwater flux into the North Atlantic is expected. In addition precipitation at high latitudes is predicted to increase under global warming as well, and thus contributing to an additional increased freshwater flux. These changes can be crucial to the production of deep water taking place in the Nordic Seas and the heat transport into this region. As this heat transport is believed to be an important contributor to the mild climate of North Europe, it is speculated that this could induce a cooling of the climate of this region.

In this thesis the impact of an increased freshwater flux on the meridional overturning circulation in a three-dimensional ideal version of the North Atlantic is investigated by performing several numerical experiments in which the freshwater flux is altered. The energetics driving the circulation are analyzed by the use of the streamfunction in depth-density coordinates, as proposed by Nycander et al. (2007). The experiments have been conducted for two different basins, one with constant depth and one including bottom topography. The modern terrain following vertical coordinate model ROMS is the model used to conduct these experiments. Traditionally, numerical studies of ocean climate have been performed by the use of geopotential vertical coordinate models. To justify our choice of model the experiments described in Marotzke (1997) and Nycander et al. (2007) were recreated by the use of ROMS. A comparison of the results shows sufficient agreement between the results from the different model types to justify its use for this purpose.

The experiments with an increased freshwater flux were consistent in predicting a weakening of the meridional overturning circulation, and thus reduced deep water production. Associated with this weakening is a decrease of surface layer temperatures at the northern boundary and a heating of the remaining ocean domain. A shutdown of the deep water production at the northern boundary and reversal of the overturning circulation is predicted for very large freshwater fluxes. Fluxes of this magnitude are however deemed highly unlikely. An interesting result is that the presence of bottom topography makes the meridional overturning circulation more resistant to changes in the freshwater flux, and only a weakening of the circulation is predicted in these experiments. The analysis of the energetics driving the circulation shows that it is chiefly thermohaline driven. The spatial resolution of the ocean basin in question along with the density intervals used to calculate this streamfunction may however be sources of error in this calculation. Parameterization of subgrid processes in the model may also affect the results.
Acknowledgments

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Chapter 1

Introduction

Due to features as the Gulf Stream, formation of deep water and a strong meridional overturning, the North Atlantic is the subject of numerous studies. The meridional overturning circulation, of which the Gulf Stream is an important part, is acknowledged as a contributory reason for the mild climate of western Europe (Ager-Wick, 2008). Deep water is formed in the Nordic Seas during wintertime, as surface water is cooled to the freezing point. This causes an increase of density at the surface and the stratification of the underlying water column to become unstable, allowing water to sink from the surface to great depths.

Global warming due to a buildup of greenhouse gases in the atmosphere causes a decay of sea ice and glaciers, particularly on the northern hemisphere. Additionally, it is predicted that global warming will lead to increased precipitation at high latitudes (IPCC, 2007). This means that the amounts of freshwater supplied to the ocean at these latitudes will increase. This may be crucial for the production of deep water. If the surface salinity is decreased to such an extent that the surface water no longer attain sufficient high density for deep water formation, it is speculated that the meridional overturning circulation will suffer a complete shutdown. As this circulation is associated with a considerable heat transport into the Nordic Seas, it is claimed that this could induce a significant cooling of the climate of western Europe even under global warming (Marotzke, 2000).

However, a complete shutdown of the meridional overturning circulation is unlikely as the meridional overturning is not driven by deep convection, but turbulent mixing which is driven by wind forcing and internal tidal waves (Wunsch and Ferrari, 2004). Thus, even if the deep water formation comes to a halt, the turbulent mixing sustains (Røed et al., 2004). It is possible that the surface currents carrying cold, low salinity water will increase in strength instead. However, if the surface layer of freshwater grows so deep that the wind fails to mix it with the saline waters beneath, a situation where the warm, saline waters flowing in from lower latitudes loses contact with the surface may arise. This resembles today's situation in the Arctic Ocean. If an ice cover of the Nordic Seas was to develop, the impact on the climate of northern Europe would be severe.

The aim of this thesis is to gain understanding of the sensitivity of the meridional overturning circulation in the North Atlantic to increased freshwater fluxes at high latitudes. Is a shutdown of the overturning probable under global warming? Will the circulation be
reversed? That is, is a situation possible where saline water sinks at the equator, flows north at depth, gradually cool and freshen, rise back to the surface and return south as a surface current of cold, low-salinity water, arise?

The thermohaline circulation has been the subject of numerous studies, both by numerical and analytical means. Sandström (1908) and Stommel (1961) were pioneers in this field, and they both pointed out the need of an external mixing source to sustain the circulation. Manabe and Stouffer (1994) provide a coupled atmosphere-ocean study in which the effect of increased atmospheric carbon dioxide concentrations on the ocean are analyzed. A substantial weakening of the overturning circulation is predicted due to increased precipitation at high latitudes. The transient as well as equilibrium responses of the thermohaline circulation to surface freshwater fluxes are investigated in Wang et al. (1999a,b). While a collapse of the overturning circulation was predicted in the transient case, the equilibrium case predicted a strengthening of the overturning circulation.

Based on numerous numerical experiments in which the freshwater flux into the North Atlantic is varied, we will investigate the effect of an increased freshwater flux at high latitudes, as a result of melting of sea ice as well as glaciers and increased precipitation, on the meridional overturning circulation is investigated. To analyze the driving forces of the ocean circulation in our experiments, the streamfunction in depth-density coordinates, as advocated by Nycander et al. (2007) is used.

Previous numerical studies of the meridional overturning circulation have been carried out by the use of geopotential coordinate models whereas we will make use of a modern terrain following vertical coordinate model (ROMS). To satisfy ourselves that ROMS is suitable for this purpose we have replicated the experiments performed by Marotzke (1997) and Nycander et al. (2007) and a comparison of the results has been made.

The thesis is organized as follows: In Chapter 2 a brief description of the thermohaline circulation and a way of analyzing its energetics is given. The model used to conduct the
experiments this thesis is based on is described in Chapter 3. In Chapter 4 a verification of this model is given, along with descriptions of the different configurations. Chapter 5 provides the results from the experiments in which the freshwater flux was increased. These results are discussed in Chapter 6. Finally, in Chapter 7 a summary and a conclusion of the thesis is presented.
Chapter 2

Theory

2.1 Thermohaline circulation

The ocean circulation can be defined as the circulation of its mass. This includes circulation of all other sea water properties, such as heat, salt and oxygen. The circulation of a sea water property itself may differ substantially from the circulation of other properties. Traditionally, the ocean circulation is divided into two parts, the wind driven and the thermohaline circulation (Pickard and Emery, 1990).

Thermohaline circulation is a term for the fraction of the ocean circulation that is driven by density variations. As sea water density is determined chiefly by temperature and salinity, it can be regarded as a circulation of these properties. However, salinity and temperature have quite different surface boundary conditions and it follows that the circulation of temperature and salinity should be investigated separately. It should be noted that the atmospheric winds have great influence on these boundary conditions (Wunsch, 2002).

In this thesis wind forcing is omitted, and the circulation in question can thus be regarded as purely thermohaline. Here, we will be particularly concerned about the meridional overturning circulation which is obtained by integrating the meridional ocean flow component below a surface of constant depth, zonally (Nurser and Lee, 2004a),

$$\psi(y, z) = \int \int v \, dx \, dz.$$  \hspace{1cm} (2.1)

Consider an ocean with flat bottom and straight coasts, extending from the equator to the North Pole. For simplicity salinity is omitted and the temperature is assumed constant. At a time $t$ atmospheric heat forcing is turned on. The surface layer is heated at the equator, causing a decrease in the density here. Due to thermal expansion, the surface will be slightly elevated as well. At the northern boundary, our ocean loses heat to the atmosphere, causing the surface waters to cool down, thus increasing the density. This increased density causes the water column to become statically unstable and the cold, dense surface waters start to sink. Due to the surface elevation at the equator, a northward motion in the surface layer will be initiated. At the northern end the dense waters having reached the bottom will start to spread out, resulting in a southward flow along the ocean floor. As the dense water...
flow towards the equator it will mix with the overlying waters and gradually its density will
decrease allowing the water to re-emerge to the surface. The warm water in the surface
will be cooled on its way north until it is dense enough to sink and thus, closing the circuit
(See Figure 2.1). A circulation driven by these mechanisms is said to be a temperature
dominated circulation, and will in this thesis be defined as positive overturning.

Now, consider the same ocean, but replace the surface boundary condition with the follow-
ing: Due to increased precipitation and melting of sea ice as well as the glaciers of Northern
Europe, a freshwater flux is added at high latitudes, while at the equator evaporation leads
to more saline waters. The result is an increase of density in the southern end and a de-
crease in the northern end. The light surface waters of the north will flow towards the
equator while the dense water there sinks to the bottom and spread out towards the north
(See Figure 2.2). This circulation is a salinity dominated one, defined here as negative
overturning.

Unlike the two-dimensional ideal oceans discussed above, both temperature and salinity
forcing must be applied simultaneously when the two-dimensional meridional overturning
circulation of a real three-dimensional ocean is to be investigated. As salinity and tem-
perature have opposing effects on density, the sign of the overturning circulation depends
upon which property dominates the density. In this thesis we seek to answer the following
questions: How does an increased freshwater flux affect the overturning circulation and the temperature distribution of the ocean? Is it possible that this circulation will suffer a complete shutdown as a result of global warming? Is a salinity dominated overturning circulation probable in the future?

Stommel (1961) proposed a simplified mathematical approach for studying the dependence of the meridional overturning circulation on temperature and salinity. His model is shown in Figure 2.3. Additionally, a mixing device was included. The changes in \( T_1, T_2, S_1 \) and \( S_2 \) can be expressed as

\[
\begin{align*}
\partial_t T_1 &= -|q|T_1 + |q|T_2 + H_T, \\
\partial_t T_2 &= |q|T_1 - |q|T_2 - H_T, \\
\partial_t S_1 &= -|q|S_1 + |q|S_2 + H_S, \\
\partial_t S_2 &= |q|S_1 - |q|S_2 - H_S,
\end{align*}
\]

respectively. Here \( H_T \) represents the heat flux, while \( H_S \) is the freshwater flux. It follows that \( q \) is proportional to the density difference,

\[
q = \kappa (\rho_2 - \rho_1) = \kappa [\alpha (T_2 - T_1) - \beta (S_2 - S_1)] = \kappa (\alpha \Delta T - \beta \Delta S),
\]

where \( \Delta T \) and \( \Delta S \) represent the temperature and salinity difference between low and high latitudes, respectively. \( \kappa \) is a hydraulic constant, \( \alpha \) and \( \beta \) are the expansion coefficients for temperature and salinity, respectively. This means that the sign of \( q \) depends on whether \( \alpha \Delta T \) is greater than \( \beta \Delta S \) or not. If \( q > 0 \) the circulation is temperature dominated (positive overturning) and the salinity difference act as a restrainer. For \( q < 0 \) the situation is reversed. Subtracting (2.5) from (2.4) leads to

\[
\partial_t \Delta S = 2|q|\Delta S - 2H_S.
\]

Keeping \( \Delta T \) constant and solving (2.7) for a steady state for the salinity difference \( \Delta \hat{S} \) leads to

\[
H_S = \kappa |\alpha \Delta \hat{T} - \beta \Delta \hat{S}|\Delta \hat{S}.
\]

Assuming a linear equation of state, this equation has two stable solutions, one for \( q > 0 \) and \( \alpha \Delta \hat{T} > \beta \Delta \hat{S} \) and the second for \( q < 0 \) and \( \alpha \Delta \hat{T} < \beta \Delta \hat{S} \). The first solution results in a temperature dominated circulation with sinking at high latitudes, while the second gives a salinity dominated circulation with sinking at low latitudes. Allowing a nonlinear equation of state however, creates multiple equilibria (Marotzke, 2000).
2.2 The streamfunction in depth-density coordinates

Whether the ocean circulation is driven by density gradients or mechanically forced has been debated for a long time. Sandström (1908) performed several laboratory experiments and concluded that a thermally forced circulation could only occur if the heating takes place at a higher pressure than the cooling. In the ocean, however, both cooling and heating takes place at the surface. The solar radiation penetrates, at the most, the upper 100 meters (Pickard and Emery, 1990) and the resulting circulation should consequently be confined to the region above this depth. During the past century his theories have gained support and several papers have concluded that the circulation must be mechanically forced (Marotzke, 2000; Wunsch and Ferrari, 2004; Defant, 1961; Jeffreys, 1925).

When considering the energetics of the ocean circulation, it is obvious that the deep convection of the north releases potential energy and can thus be considered as an energy sink. Having reached the bottom the water flows southward and in order to close the circuit, the water must rise back towards the surface. As the ocean is nearly everywhere stable stratified, this means crossing the stratification. In order to achieve this the presence of turbulent mixing is necessary. A circulation as described in Section 2.1 should in other words not be possible without providing a energy source for the mixing. The energy sources that sustains this mixing is discussed in Wunsch and Ferrari (2004). First and foremost wind forcing is the main contributor in driving turbulent mixing. The wind contributes both directly by generating surface waves and indirectly by redistributing mass in the upper layers. This leads to the generation of internal waves (Sandström, 1908), which breaks when they reach the continental shelfs. Secondly, internal tidal waves also contributes to the turbulent mixing in the abyss, but wind forcing remains the primary energy source. Other sources contribute to the turbulent mixing as well, but in a much lesser extent.

To analyze the energetics of the ocean circulation Nycander et al. (2007) advocated the use of a streamfunction in depth-density coordinates, introduced by Nurser and Lee (2004a). This streamfunction can be computed by calculating the vertical (upward) transport across levels of constant depth between isopycnals. I have chosen to use the equation given by Nurser and Lee (2004a):

\[
\psi^* (\theta_a, z) = \int \int_{(x,y)\in \Omega(x,y,z) \leq \theta_a} w \, dx \, dy, \quad (2.9)
\]

where \( \theta \) represent potential temperature and \( \theta_a \) is a specific temperature. It should be noted that as density depends upon both temperature and salinity, potential temperature in (2.9) have been interchanged with potential density in the calculations presented later in this thesis. In ocean models small scale mixing is parameterized as diffusion, this means that a circulation that is forced by small scale mixing will appear as driven by density gradients (thermohaline driven). Notice that in order for the definition of the streamfunction to be consistent, the flow in question must be in a steady state, at least in a statistical sense (quasi steady state). As pressure is linearly related to depth by the hydrostatic relation

\[
\partial_z p = -\rho g, \quad (2.10)
\]

and the specific volume is given as \( \sigma = 1/\rho \), this streamfunction in effect displays the overturning circulation in a \( pV \) diagram which is the traditional way to describe a thermal
2.2 The streamfunction in depth-density coordinates

Figure 2.4: An overturning circulation cell displayed by the streamfunction in depth-density coordinates. The clockwise motion indicate that the circulation is driven by buoyancy forcing. In the case of salinity dominated overturning the direction of the flow will be conserved, but the position of the boundaries will be interchanged. The figure is copied from Nurser and Lee (2004a).

cycle within the field of thermodynamics. The direction of the flow in an overturning cell directly shows whether the cell is driven mechanically (counter-clockwise) or by density differences (clockwise). A schematic example of this streamfunction is shown in Figure 2.4. We observe that in this circulation cell dense water sinks at the northern boundary, and as it rises towards the surface at low latitudes the density decreases gradually. A circulation cell with this appearance is termed a thermohaline driven cell. The circulation pattern in this figure coincides with a temperature dominated circulation. In the case of a salinity dominated circulation the general pattern will persist, but the northern boundary will switch location with the southern boundary, but it remains to be a thermohaline driven circulation. In contrast a mechanically driven cell is characterized by light water sinking, gaining higher density on its way downward. As the dense water is raised back towards the surface, its density gradually decreases. It is obvious that the latter circulation pattern needs an input of energy in order to sustain. For further details, see Nycander et al. (2007); Nurser and Lee (2004a,b).
Chapter 3

Model Description

The aim of this thesis is to investigate the meridional overturning circulation of the North Atlantic and its sensitivity to an increased freshwater flux. In order to achieve this we will use the recently developed, community model ROMS.

3.1 General description

The Regional Ocean Modeling System (ROMS) is a free surface, terrain-following primitive equations ocean model which may be applied to solve a wide range of oceanic problems. During the past years several versions of the model have been released. The version applied in this thesis is ROMS 3.0 which was released April, 2007. The source code is written in Fortran 90/95 and uses C-preprocessing to activate physical and numerical options chosen by the user.

In order to increase the model’s efficiency, a split-explicit time stepping scheme for the barotropic (fast) and baroclinic (slow) modes is used (IMCS). This means that a finite number of barotropic steps are carried out within a baroclinic time step in order to resolve the external gravity waves.

ROMS uses a stretched terrain-following s-grid, which allows better resolution in layers of particular interest such as the bottom boundary layer and the thermocline. The relationship between the true vertical coordinate $z$ and $s$ is given by

$$z = \zeta + \left(1 + \frac{\zeta}{h}\right)[h_c s + (h - h_c)C(s)]. \quad -1 \leq s \leq 0,$$

(3.1)

Here, $\zeta(x,y,t)$ is the free surface deviation, $h(x,y)$ represents the bottom topography, $h_c$ is the critical depth and $C(s)$ is given by

$$C(s) = (1 - b)\frac{\sinh(\theta s)}{\sinh \theta} + b\frac{\tanh \left[\theta(s + \frac{1}{2})\right] - \tanh(\frac{\theta}{2})}{2 \tanh(\frac{\theta}{2})},$$

(3.2)

where $\theta$ and $b$ are the surface and bottom control parameters, whose ranges are $0 \leq \theta \leq 20$ and $0 \leq b \leq 1$ (Song and Haidvogel, 1994). The new coordinate $s$ ranges from $-1$ at the

1http://www.myroms.org
bottom to 0 at the surface. This coordinate ensures that the highest resolution is near the surface, as long as \( \theta \) is chosen correctly. The vertical grid is staggered, and is shown in Figure 3.1a. The use of a \( \sigma \)-coordinate introduces pressure gradient errors, due to the splitting of the horizontal pressure gradient into a component along the \( \sigma \)-surface and a correction term (Griffies, 2004)\(^2\). ROMS, however, is designed to minimize these errors (IMCS).

In the horizontal, it is possible to use both Cartesian and spherical coordinates. The primitive equations are evaluated on a staggered Arakawa C-grid (Figure 3.1b), where \( \rho \), \( u \) and \( v \) points all have their distinct locations. The zonal coordinate is denoted \( \xi \) and the meridional coordinate \( \eta \). ROMS is a Boussinesq approximation model, which means that density variations are neglected except when multiplying with gravity. In other words, the density and mass of a fluid parcel may change, but its volume remain constant (Griffies, 2004).

### 3.2 The governing equations

Like all ocean models ROMS is based on the momentum equations, the continuity equation, an equation of state and tracer equations. Here, only the active tracers (tracers that effect the ocean density) are included. A summary of the main variables in ROMS is shown in Table 3.1. The horizontal momentum equation is

\[
\partial_t u + \nabla_H \cdot (\rho u v) + f k \times u = -\frac{1}{\rho_0} \nabla_H p + F + D, \tag{3.3}
\]

\[
(3.4)
\]

where \( u = u_i + v_j \), \( v = u_i + v_j + w_k \) and \( \nabla_H = i \partial_x + j \partial_y \), \( D \) and \( F \) represent the horizontal diffusive and forcing terms, respectively. \( f \) is the Coriolis parameter and \( \rho_0 \) is a constant reference density. The vertical momentum equation is simply the hydrostatic relation,

\[
\partial_z p = -\rho g, \tag{3.5}
\]

\(^2\)http://www.myroms.org
where \( g \) is the gravitational acceleration. The continuity equation is due to the Boussinesq approximation simply

\[
\nabla_H \cdot \mathbf{u} + \partial_z w = 0, \tag{3.6}
\]

and the equation of state as

\[
\rho = \rho(T, S, p). \tag{3.7}
\]

Additionally we have two tracer equations, one for potential temperature and one for salinity:

\[
\begin{align*}
\partial_t T + \mathbf{v} \cdot \nabla T &= F_T + D_T, \tag{3.8} \\
\partial_t S + \mathbf{v} \cdot \nabla S &= F_S + D_S. \tag{3.9}
\end{align*}
\]

Here, \( D \) represents the diffusive terms while \( F \) represents the forcing terms. In order to be able to use the terrain-following vertical coordinate, the coordinates have to be transformed according to

\[
\begin{align*}
t' &= t, \\
x' &= x, \\
y' &= y, \\
s &= s(x, y, z),
\end{align*}
\]

Applying these transformations to the partial derivative operators yield

\[
\begin{align*}
\nabla_H &= \nabla_s - \nabla_s z \left( \frac{1}{H_z} \right) \frac{\partial}{\partial s}, \\
\frac{\partial}{\partial z} &= \left( \frac{\partial s}{\partial z} \right) \frac{\partial}{\partial s} = \frac{1}{H_z} \frac{\partial}{\partial s}.
\end{align*}
\]

Here, \( H_z = \partial_s z \) is the specific thickness and \( \nabla_s = i \partial_x + j \partial_y \). Applying these to (3.3) - (3.9) results in the slightly more complicated equations (primes are dropped for clarity):

**Momentum:**

\[
\begin{align*}
\partial_t (H_Z \mathbf{u}) + \nabla_s \cdot (\mathbf{u} \mathbf{u}) + \partial_s (\mathbf{u} \Omega) + f \mathbf{k} \times \mathbf{u} &= -\frac{1}{\rho_0} \nabla_H p - \frac{g \rho}{\rho_0} \nabla_s z - g \nabla_s \zeta + \mathbf{F} + \mathbf{D} \tag{3.12} \\
\partial_s p &= -g \rho H_z. \tag{3.13}
\end{align*}
\]

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description and unit</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \zeta )</td>
<td>free surface deviation [m]</td>
<td>( (\eta_\rho, \xi_\rho) )</td>
</tr>
<tr>
<td>( \mathbf{u} )</td>
<td>vertically integrated u-momentum [m/s]</td>
<td>( (\eta_\mathbf{u}, \xi_\mathbf{u}) )</td>
</tr>
<tr>
<td>( \mathbf{v} )</td>
<td>vertically integrated v-momentum [m/s]</td>
<td>( (\eta_\mathbf{v}, \xi_\mathbf{v}) )</td>
</tr>
<tr>
<td>( u )</td>
<td>u-momentum component [m/s]</td>
<td>( (s_\rho, \eta_\mathbf{u}, \xi_\mathbf{u}) )</td>
</tr>
<tr>
<td>( v )</td>
<td>v-momentum component [m/s]</td>
<td>( (s_\rho, \eta_\mathbf{v}, \xi_\mathbf{v}) )</td>
</tr>
<tr>
<td>( \Omega )</td>
<td>s-coordinate vertical momentum [m³/s]</td>
<td>( (s_w, \eta_\mathbf{w}, \xi_\mathbf{w}) )</td>
</tr>
<tr>
<td>( w )</td>
<td>vertical momentum component [m/s]</td>
<td>( (s_\mathbf{w}, \eta_\mathbf{w}, \xi_\mathbf{w}) )</td>
</tr>
<tr>
<td>( T )</td>
<td>potential temperature [°C]</td>
<td>( (s_\rho, \eta_\mathbf{T}, \xi_\mathbf{T}) )</td>
</tr>
<tr>
<td>( S )</td>
<td>salinity [none]</td>
<td>( (s_\rho, \eta_\mathbf{T}, \xi_\mathbf{T}) )</td>
</tr>
<tr>
<td>( \rho )</td>
<td>in situ density anomaly [kg/m³]</td>
<td>( (s_\rho, \eta_\mathbf{T}, \xi_\mathbf{T}) )</td>
</tr>
</tbody>
</table>

Table 3.1: Main variables in ROMS.
Continuity:
\[ \partial_t H_z + \nabla_H \cdot (H_z u) + \partial_s (H_z \Omega) = 0. \]  
Equation of state:
\[ \rho = \rho(T, S, p). \]  
Tracer equations:
\[ \partial_t (H_z T) + \nabla_s \cdot (H_z T u) + \partial_s (H_z T \Omega) = F_T + D_T, \]  
\[ \partial_t (H_z S) + \nabla_s \cdot (H_z S u) + \partial_s (H_z S \Omega) = F_S + D_S. \]

Here, \( \Omega \) is the velocity component perpendicular to the \( s \)-surfaces, given as
\[ \Omega(x, y, s, t) = \frac{1}{H_z} \left[ w - (1 + s) \partial_t \zeta - u \partial_x z - v \partial_y z \right]. \]

The velocity vector \( \mathbf{v} \) now have the components \( u, v \) and \( \Omega \). The true vertical velocity component \( w \) is related to the generalized vertical velocity \( \Omega \) by
\[ w = \partial_t z + \partial_x z + \partial_y z + \Omega H_z. \]

In order to be able to use spherical coordinates the new curvilinear horizontal coordinates \( \xi(x, y) \) and \( \eta(x, y) \) are introduced. The relationship between the horizontal arc length and the differential distance is given by
\[ (ds)_\xi = \frac{1}{m} d\xi, \]  
\[ (ds)_\eta = \frac{1}{n} d\eta. \]

Here, \( m(\xi, \eta) \) and \( n(\xi, \eta) \) relate the differential distances \( (\Delta \xi, \Delta \eta) \) to the physical arc lengths. This formulation of the curvilinear coordinates transform back into Cartesian coordinates by setting \( m = n = \text{constant} \) (Haidvogel et al., 2000). (3.12) - (3.17) can now be re-written,
\[ \partial_t \left( \frac{H_z u}{mn} \right) + \partial_\xi \left( \frac{H_z u^2}{n} \right) + \partial_\eta \left( \frac{H_z uv}{m} \right) + \partial_s \left( \frac{H_z u \Omega}{mn} \right) - \left[ \left( \frac{f}{mn} + v \partial_\xi \left( \frac{1}{n} \right) \right) \right] H_z v = - \left( \frac{H_z}{n} \right) \left( \frac{1}{\rho_0} \partial_\xi p + \frac{g \rho_0}{\rho_0} \partial_\xi z + g \partial_\xi \zeta \right) + \frac{H_z}{mn} (F_u + D_u), \]  
\[ \partial_t \left( \frac{H_z v}{mn} \right) + \partial_\xi \left( \frac{H_z v^2}{n} \right) + \partial_\eta \left( \frac{H_z v^2}{m} \right) + \partial_s \left( \frac{H_z v \Omega}{mn} \right) - \left[ \left( \frac{f}{mn} + v \partial_\xi \left( \frac{1}{n} \right) \right) \right] H_z u = - \left( \frac{H_z}{m} \right) \left( \frac{1}{\rho_0} \partial_\eta p + \frac{g \rho_0}{\rho_0} \partial_\eta z + g \partial_\eta \zeta \right) + \frac{H_z}{mn} (F_v + D_v), \]  
\[ \partial_t \left( \frac{\zeta}{mn} \right) + \partial_\xi \left( \frac{H_z u}{n} \right) + \partial_\eta \left( \frac{H_z v}{m} \right) + \partial_s \left( \frac{H_z \Omega}{mn} \right) = 0, \]  
\[ \partial_s p = -g \rho H_z, \]  
\[ \rho = \rho(T, S, p), \]
\[ \begin{align*}
\partial_t \left( \frac{H_zT}{mn} \right) + \partial_\xi \left( \frac{H_zuT}{n} \right) + \partial_\eta \left( \frac{H_zvT}{m} \right) + \partial_s \left( \frac{H_z\Omega T}{mn} \right) &= \frac{H_z}{mn} (F_T + D_T), \\
\partial_t \left( \frac{H_zS}{mn} \right) + \partial_\xi \left( \frac{H_zuS}{n} \right) + \partial_\eta \left( \frac{H_zvS}{m} \right) + \partial_s \left( \frac{H_z\Omega S}{mn} \right) &= \frac{H_z}{mn} (F_S + D_S).
\end{align*} \tag{3.27, 3.28} \]

A thorough description of the model is given in Song and Haidvogel (1994); Haidvogel et al. (2000, 2008).
Chapter 4
Verification

The simplest and most common choice for vertical coordinate is $z$, the vertical distance from the surface. Ocean models applying this coordinate are often termed geopotential coordinate models and are widely used in studies of ocean climate. Bottom topography is difficult to represent with this choice of coordinate and thus bottom boundary layer processes are difficult to parameterize. The free surface also yields difficulties, as there has to be a restrain on the surface deviation in order to avoid vanishing surface grid cells. This is problematic especially in regions where the tidal fluctuations are large (Griffies, 2004). The advantage of this coordinate is that the numerical discretization of the governing equations can be quite simple with this choice of vertical coordinate, and it is straightforward to obtain sufficient resolution of the surface mixed layer (Griffies, 2004).

In order to better represent the bottom topography and the physical processes associated with the bottom boundary layer, the terrain following vertical coordinate $\sigma$, given as

$$\sigma = \frac{z - \eta}{H + \eta},$$

where $\eta$ represents the surface displacement and $H$ the ocean bottom, was introduced. Models applying this vertical coordinate are particularly popular in studies of coastal phenomena. This coordinate have some disadvantages as well. The resolution of the surface mixed layer can be quite coarse in regions off the continental shelves. Additionally, as discussed in Chapter 3, the pressure gradient error is introduced. In ROMS both of the issues above are minimized by use of the modified terrain following coordinate and sophisticated numerics.

In past numerical studies of the overturning circulation mostly geopotential coordinate models have been used. Thus, it is interesting to see how well the results from ROMS correspond to earlier studies. The aim of this chapter is to verify that this model indeed can be used for studies of this kind. A comparison of the ROMS results with earlier experiments applying similar configurations, but using a geopotential coordinate model, are therefore given in this chapter.
Figure 4.1: Forcing profiles of the simulations. The solid line represents both temperature and salinity forcing for the Marotzke (1997) configuration, with the temperature scale shown on the left-hand vertical axis and salinity along the right-hand axis. The dashed line represents the temperature forcing for the Nycander et al. (2007) configuration.

4.1 Marotzke

As mentioned in the introductory chapter, similar experiments were first conducted by Marotzke (1997). He used The Geophysical Fluid Dynamics Laboratory (GFDL) primitive equation GCM, a geopotential coordinate model, to perform his simulations.

The ocean basin extends from the equator to 64°N and has a constant depth of 4500 meters. In the horizontal, spherical coordinates are applied. The resolution is quite coarse, 3.75° zonally and 4.0° meridionally which gives 16 x 16 points horizontally. There are 15 vertical levels, originally the resolution varied from 50 meters near the surface to 500 meters near the bottom. Horizontal and vertical viscosity coefficients are 2.5 x 10^5 and 10^-2 m^2s^-1, respectively. Horizontal diffusivity is 10 m^2s^-1. Static instability is removed by an convective scheme that increases the vertical diffusivity to 1 m^2s^-1. The vertical diffusivity under stable conditions is zero everywhere except in the columns adjacent to the lateral boundaries where a vertical mixing coefficient of 5 x 10^-4 is applied. This is done to mimic the effect of sloping lateral boundaries where mixing is believed to be strong, due to breaking of internal waves. At the surface both temperature and salinity are restored towards the profiles

\[ T = 27^\circ C + \frac{\Delta T (\cos(\frac{\pi \varphi}{64}) - 1)}{2}, \]
\[ S = 36 + \frac{\Delta S (\cos(\frac{\pi \varphi}{64}) - 1)}{2}, \]

where \( \varphi \) is the latitude, \( \Delta T = 27^\circ C \) and \( \Delta S = 1.5 \) (See Figure 4.1). The restoring time-scale is 30 days, which is so rapid that the sea surface density almost is prescribed. The density is calculated by a nonlinear equation of state and a no-slip boundary condition is applied at the lateral boundaries.

In our verification experiment we have been faithful to the configuration described above. Our vertical resolution is however different, it varies from 36 meters at the surface to approximately 750 meters at the bottom. In order to achieve numerical stability the baroclinic time-step had to be 1.6 hours for our configuration. Marotzke (1997) provides
Figure 4.2: a) The streamfunction from the Marotzke (1997) experiment. The figure is copied from Marotzke (1997). b) the streamfunction from the ROMS run. The red shading indicates positive overturning. Contour interval is 2 Sv for both figures, but in b) the 1 Sv and 0.5 Sv contours have been included as well. Note the different scaling of the vertical axes.

no information on the time-step length, but as the configuration is nearly identical to the one used earlier in Marotzke and Willebrand (1991) and Weaver et al. (1993), in which the time-step used is two hours for the momentum equations, it is probable that this is the time-step used also in Marotzke (1997).

The experiment was started from an ocean at rest with constant temperature and salinity. The initial temperature was 5°C and the salinity was 33. The model was run for 2000 successive years in order to ensure a steady state. As the depth of the basin is constant, the $s$-surfaces in this case are horizontally aligned and thus ROMS essentially act as a geopotential coordinate model for this case, but with a slightly different vertical distribution of $z$-levels. The results from the ROMS simulation is given below.

To ease the comparison the meridional overturning streamfunction for both the Marotzke run and the ROMS run is shown in Figure 4.2. Both figures consist of a single, positive overturning cell of the same order of magnitude (Maximum of 19.2 Sv in the Marotzke run compared to 19.3 Sv in the ROMS run). The maximum of the ROMS run is located further north and is approximately 1000 meters shallower than in the Marotzke run. This causes the northward flow at the western boundary to increase monotonically with latitude at the first model level (Figure 4.6), whereas it peaks mid-basin in the Marotzke run (not shown). The horizontal flow in the abyss is correspondingly weaker in the ROMS run. The coarser resolution in the abyss of our experiment could be the cause of both the vertical shift of the maximum transport as well as the weaker horizontal flow at depth.

Temperature profiles along the western boundary for both models are shown in Figure 4.3. The figures show a striking resemblance, with strong stratification which decreases with latitude and depth. At the eastern boundary this resemblance is less obvious (Figure 4.4). While the geopotential coordinate model predicts deep convection and neutral stability increasing with latitude, ROMS predicts stable stratification at all latitudes below a depth of approximately 700 meters. In the layers above this, however, the temperature profiles show the same trends as in the original run. Also the downward slope of the 5, 4 and 3°C
Figure 4.3: a) show the temperature along the western boundary from the geopotential coordinate model, copied from Marotzke (1997). b) shows the same from the ROMS run. Contour interval is 1°C. Note the different scaling of the vertical axes.

contours show a similar tendency to that of Marotzke (1997). But when the structure of the temperature profiles throughout the basin is examined, we find that the deepest level of neutral stability is found at approximately 24° east of the western boundary (Figure 4.5). This suggest that the three dimensional flow pattern of the two runs differ from each other. Figure 4.6 show the horizontal velocity at the first model level, located at 17 meters depth, superimposed on the horizontal temperature distribution. A striking feature is the flow pattern at the northern boundary. As the intense northward flow at the western boundary approaches high latitudes it is deflected towards the east. When it reaches the eastern boundary it turns back around and return west to the north of this flow. Marotzke (1997) also show an intensified current at the western boundary which deflects to the east, but

Figure 4.4: a) show the temperature along the eastern boundary from the geopotential coordinate model, copied from Marotzke (1997). b) shows the same from the ROMS run. Contour interval is 1°C. Note the stable stratification found at all depths at the northern boundary in panel b). Note the different scaling of the vertical axes.
4.1 Marotzke

Figure 4.5: Temperature profile along the longitude of 24° from the ROMS run. Contour interval is 1°C. Note the vertical alignment of the 3°C-contour at the northern boundary, reaching down to a depth of 1500 meters.

Figure 4.6: Horizontal temperature distribution on the first model level (17 meters) and horizontal velocity at the same depth. Contour interval is 1°C, reference vector represents a velocity of 5 cm/s (blue arrow at the top) Note the strong current at the western boundary and the eastward jet at high latitudes.

Although there are some differences between the results from the two models, these are not of crucial importance. These differences could be caused by different discretization of the governing equations, different parameterization of subgrid scale processes as well as the use of more sophisticated numerics in ROMS. The different vertical resolution may also contribute to these differences. Based on this experiment there is no evidence which suggests that ROMS is unfit to model the meridional overturning circulation.
Figure 4.7: The results from Nycander et al. (2007). a) Temperature field of a mid-basin cross-section and b) meridional streamfunction. The transport is counted positive when the circulation is anticlockwise. The figure was copied from Nycander et al. (2007). Note the shallow overturning compared to the constant depth configuration.

4.2 Nycander

In this section we show results from an experiment similar to the experiment described in Nycander et al. (2007). They applied the Massachusetts Institute of Technology general circulation model (MITgcm), which is a geopotential coordinate model with more sophisticated numerics than the model used in Marotzke (1997), to conduct their experiments. Otherwise the main difference between the Marotzke (1997) and Nycander et al. (2007) experiments is the presence of bottom topography in th latter.

The domain is $6000 \times 6000$ km and situated at the northern hemisphere. Cartesian coordinates on an $f$ plane are applied, with $f = 10^{-4}$ s$^{-1}$. The horizontal resolution is 100 km in both $x$- and $y$-direction. The depth of the ocean increases linearly from the coast until it reaches its maximum depth of 3000 meters at 2000 km from the shore. There are 25 levels in the vertical, with a resolution ranging from 50 meters at the surface to 200 meters at the bottom. Horizontal and vertical viscosities are 5000 and $10^{-3}$ m$^2$s$^{-1}$, respectively. The vertical diffusivity is $10^{-4}$ m$^2$s$^{-1}$. To remove static instability a convective scheme that increases the vertical diffusivity to 1 m$^2$s$^{-1}$ when the column is unstable stratified, is parameterized. The horizontal mixing was represented by the Gent-McWilliams scheme, applying an isopycnal diffusivity of $10^3$ m$^2$s$^{-1}$. The boundary condition on the tangential velocity at the lower boundaries is no-slip. At the surface the temperature is restored towards the profile

$$T = \frac{\Delta T [\cos \left( \frac{2\pi}{L} \right) + 1]}{2},$$

where $\Delta T = 20^\circ$C and $L$ is the basin length (See Figure 4.1). The restoring time-scale is 12.5 days, so the surface density is nearly prescribed. The density depends linearly on temperature.
Due to time limitations, some of our experiments applied a coarser horizontal resolution, the grid spacing was increased to 200 km in these runs. Additionally, the bathymetry had to be changed in order to avoid a collapse of the $s$-surfaces. The depth all points closer to the shore than $1/2\Delta x$ (or $1/2\Delta y$) is set constant and equal to the depth that would have been reached here if we allowed the depth to behave as described above. Hence, the minimum depth for a horizontal resolution of 100 km is 75 meters, and for 200 km this depth is 150 meters. Due to the sloping bottom, the vertical resolution in the ROMS configuration varies substantially. It ranges from 6 meters at the surface in the columns adjacent to the shores to 325 meters at the bottom over the maximum depth. ROMS currently do not provide an option for Gent-McWilliams mixing, so in our experiments the horizontal diffusivity is simply $10 \text{ m}^2\text{s}^{-1}$. However, momentum is mixed along surfaces of constant geopotential while tracers are mixed along surfaces of constant density. It should also be noted that while the experiment in Nycander et al. (2007) was conducted on an $f$ plane, our experiments have been performed on a $\beta$ plane with $\beta = 2.0 \cdot 10^{-11}$. This may cause some differences.

No information of the time-step was available from Nycander et al. (2007), but in order for the our configuration to be stable, the baroclinic time-step had to be 36 minutes in the high resolution ROMS run. The run in which the resolution was reduced to 200 km horizontally required a baroclinic time-step length of 1 hour and 20 minutes. The temperature field of a mid-basin crosssection and the meridional overturning streamfunction from Nycander et al. (2007) are shown in Figure 4.7 a) and b), respectively. It should be noted that Nycander et al. (2007) defines the transport as positive when the circulation is anti-clockwise, contrary to the definition given in Section 2.1 in this thesis. To ease the comparison Nycander’s definition is applied in the verification experiments described next in order to make the comparison straight forward. The same initial conditions as described in Section 4.1 were applied, and the model was integrated for 2000 successive years.

Two verification runs were conducted, one with the same horizontal resolution as in Nycander et al. (2007), and one where this resolution was decreased to 200 km $\times$ 200 km. There is a major difference in the configuration of these runs; for the fine resolution case a non-linear equation of state had to be applied in order to achieve numerical stability, while a linear equation of state was used for the coarse resolution. Results from both experiments are shown in Figure 4.8. The temperature field of our high resolution run (Figure 4.8a) shows a striking resemblance to the Nycander run, with strong surface stratification which decreases when moving northwards. In the northern end there are evidence of neutral stratification, indicating sinking of cold water. The low resolution run also displays strong stratification (Figure 4.8c), but the slope of the temperature contours is much steeper in both the Nycander and high resolution run. However, indication of neutral stratification can be found in the northern end of the basin in this run as well. These facts lead to lower temperatures in the abyss for the low resolution case. The meridional overturning circulation consists of a single thermally driven cell in the high resolution case. Within this cell there are two separate cells, one in the upper 1000 meters and a weaker cell stretching from 1500 to 2500 meters depth. The highest transport value estimated is $2.5 \text{ Sv}$. The transport in the northern part of the basin is considerably weaker than in the Nycander run, where a single thermally driven circulation cell, confined to the upper 500 meters, cover the whole basin length. The meridional overturning in the low resolution run yield a more chaotic picture. A thermally driven circulation cell of maximum transport $3.7 \text{ Sv}$.
is the most dominating feature. This cell covers the basin from south to north, and is confined to the upper 750 meters, approximately. The difference in vertical extent of the main circulation cell in the two verification runs can be explained by the temperature field. The depth of the thermocline is shallower in the run where the coarse resolution is applied, and the depth of the overturning circulation depend on this depth. Beneath this cell at depths between 1000 and 1500 meters, another weaker thermally driven cell can be found. There are several small cells where the circulation is anticlockwise as well. The magnitude of the transport in both verification runs agree with the transport predicted by the Nykander run. The flow pattern is slightly different for all three runs although they all show the same trends. It should be noted that the fact that adding bottom topography leads to a shallower overturning is well reproduced by ROMS, which supports the use of terrain-following vertical coordinate models for ocean climate simulations.

Figure 4.9 shows the overturning streamfunction as a function of depth and potential dens-
Figure 4.9: The overturning streamfunction as a function of depth and potential density. The transport is counted positive (red) in anticlockwise cells and negative (blue) otherwise. a) From the Nycander run. Potential temperature is used in stead of density. The figure is copied from Nycander et al. (2007). b) From the high resolution verification run. c) From the coarse resolution verification run.
Figure 4.10: Overturning streamfunction in depth-density coordinates from the OCCAM model. The figure is copied from Nycander et al. (2007). Cells of clockwise circulation are shaded blue, while anti-clockwise circulation cells are shaded red. Transport is given in Sverdrups. Note the deep narrow thermohaline driven cell involving the densest water.
4.2 Nycander

resembles the ones presented in this thesis.

In addition to the vertical coordinate, there are other significant differences between our verification runs and the Nycander run, namely the added shelves which impose other restrictions on the flow and the horizontal diffusion scheme used. Additionally, our experiments applied the $\beta$ plane approximation, while the Nycander et al. (2007) experiment was conducted on an $f$ plane. Despite these differences the verification runs yield similar results, but the overturning circulation seems to be underestimated in the northern end of the ocean basin. Due to these results, the experiments where the freshwater flux is increased are mainly performed for the configuration described in the previous section, but results from the Nycander configuration is also presented. Despite these differences in numerics and configuration the results are overall similar. Thus we conclude that ROMS is fit for the purpose at hand, namely, to study the meridional overturning circulation of the North Atlantic.
Chapter 5

Results

As mentioned in the introductory chapter the aim of this thesis is to investigate the effect of an increased freshwater flux on the meridional overturning circulation of the North Atlantic. In Chapter 4 results from studies of this circulation under normal conditions were presented for two different basins, one with constant depth and one where topography is included. One of the key questions we raise is what happens if we increase the freshwater flux in the northern part of the basin. To possibly answer this we have done a series of experiments. The respective verification runs presented in the previous chapter are then used as a reference state in which the ocean circulation is unaffected by global warming. The main focus in this chapter is on the experiments with the flat bottom configuration, but results from the basin with bottom topography is briefly presented as well. In order to ease the comparison of the verification runs and the modified runs the colors showing zonal temperature sections and the transport of the meridional overturning streamfunction are the same in all figures.

To support the theory regarding the thermohaline circulation given in Section 2.1, an experiment in which the temperature forcing is turned off while the salinity forcing was left unaltered, was conducted. This configuration causes the ocean temperature to remain constant and equal to its initial value throughout the simulation and should hence induce a salinity dominated circulation as displayed in Figure 2.2. The resulting overturning cell is shown in Figure 5.1. The change from the default configuration (Figure 4.2b) is obvious. The overturning has shifted completely, from a positive overturning cell to a negative one as predicted. The absolute overturning strength exceeds that of the verification run, with a negative maximum value of $-30.6$ Sv compared to $19.3$ Sv. The estimated density of the water in this run is denser than in the default run due to the low temperature of $5^\circ$C, and thus, the effect of gravity as well as the fact that density no longer is pulled in different directions, causes the increased overturning strength. The result proves that salinity alone can drive a meridional overturning circulation and due to its boundary conditions the direction of the transport associated with this circulation will be opposite to that of thermally driven overturning.

Our main concern is the effect of an increased freshwater flux on the already existing overturning circulation, in which thermal forcing plays a major role. To gain understanding of the influence of a freshwater flux on the ocean circulation, several experiments with different values of $\Delta S$ in (4.3), while (4.2) remains unaltered, were conducted for the basin
Figure 5.1: The meridional overturning for the constant temperature case, contour interval 2.0 Sv, additionally the 1.0 Sv and 0.5 Sv contours have been included. Areas of negative overturning are indicated by blue shading, while areas of positive overturning are shaded red. The direction of the meridional overturning has shifted completely.

of constant depth. In Section 5.1 results from four cases representing different stages of the evolution of the overturning circulation will be presented in detail. A summary of the results from all experiments conducted with the constant depth configuration is given at the end of this section in order to provide a summary of the development. Results from the experiments with an increased freshwater flux in the bottom topography case are presented in Section 5.2.

The focus will be on the streamfunction in both depth-latitude and depth-density coordinates as well as zonal mean temperature field and temperature anomaly, calculated by subtracting the temperature field of the reference state from the temperature field of the current experiment. That it is, locations where the temperature is lower than in the reference state will yield a negative value. In order to ease the comparison of the verification run and the experiments presented, the colors showing zonal temperature sections and the transport of the meridional overturning streamfunction are the same in all figures.

5.1 Constant depth case

When the freshwater flux of the verification run is doubled, the overturning strength is further reduced, the maximum overturning strength is now reduced by 28% compared with the verification run (Figure 5.2a). As Figure 4.2b it consists of a single positive overturning cell stretching from the equator to 64°N and covering the entire basin depth. It clearly shows that the reduction is greatest at high latitudes and at depths above 2000 meters. Below this level the weakening is negligible. When examining the meridional temperature profiles, a new feature can be found in all sections in the western half of the basin; at approximately 60°N the temperature contours indicating the water colder than 5°C slope northwards indicating that warm water flowing north sinks and continues north
Figure 5.2: The three panels show results from an experiment in which $\Delta S = 3.0$ in equation 4.3. Panel a) shows the MOC (Contour interval 2.0 Sv, but the 1.0 Sv and 0.5 Sv contours have been included as well), panel b) shows the averaged temperature for the western half of the basin (Contour interval 1.0°C), and panel c) show the temperature anomaly in the upper 500 meters (Contour interval is 0.2°C for negative anomalies 0°C and 1°C for positive anomalies). In panel a) cells of positive overturning are indicated by red shading while cells of negative overturning are shaded blue. Likewise, negative anomalies in panel c) are shaded blue, and positive anomalies are shaded red. Note the negative temperature anomaly at the northern boundary.
under a layer of colder, less saline waters (Figure 5.2b). In the eastern part of the basin, this feature is completely absent, and resembles the temperature profiles found at the same locations in the verification run. The temperature anomaly is shown in Figure 5.2c. While the water at the first model level north of 44°N experience a cooling compared to the verification run, with a negative maximum value of $-0.5°C$ between 60 and 64°N, the remaining water masses are heated due to the decreased production of deep water at the northern boundary. This heating has a maximum value of 1.9°C. The cooling of the surface layer in the northern end of the basin can be explained by the temperature profiles discussed above. The increased amounts of freshwater supplied to the ocean at high latitudes causes the density to decrease to such an extent that the water at the northern boundary becomes lighter than the water flowing north (Figure 5.3). When these two water masses meet, the warm and saline water will sink and continue north below the cold water. Hence, the transport of heat into the northern most part of the basin is reduced and lead to lower temperatures. These effects can be found for experiments where $\Delta S$ is increased to a lesser extent as well, but the changes are weaker.

The strength of the overturning circulation decreases when increasing $\Delta S$ in (4.3) as expected. At $\Delta S = 2$ the overturning is weaker by 11% (Figure 5.9b) and a negative temperature anomaly can be found in the surface layer north of 40°N (not shown). The magnitude of this anomaly increases with latitude and its largest negative deviation is $-0.15°C$. The experiment in which $\Delta S = 5$ in (4.3) yield severe changes on the meridional overturning circulation (Figure 5.4a). A positive overturning cell of maximum 6 Sv still dominates, however, this cell is now confined to the upper 2000 meters. North of 56°N a negative overturning cell has appeared, covering the upper 500 meters. This cell has a negative maximum of $-1.1$ Sv. Below this negative cell, the positive cell extends further north, implying that warm, saline water originated at low latitudes, meet the cold, low salinity water of the north, sinks and continues north below the surface. The zonal mean temperature profile supports this picture (Figure 5.4b). As alluded to, the vertical restriction of the overturning cells can be explained by the density stratification of the basin. As displayed
Figure 5.4: The three panels show results from an experiment in which $\Delta S = 5.0$ in equation 4.3. Panel a) shows the MOC (Contour interval 2.0 $\text{Sv}$, but 1.0 $\text{Sv}$ and 0.5 $\text{Sv}$ contours (both positive and negative) have been included as well), panel b) shows the zonal mean temperature (Contour interval 1.0°C), and panel c) show the temperature anomaly in the upper 500 meters (Contour interval is 0.2°C for negative anomalies and 1°C for positive anomalies). In panel a) cells of positive overturning are indicated by red shading while cells of negative overturning are shaded blue. Likewise, negative anomalies in panel c) are indicated by blue shading and positive anomalies with red shading. Note the negative temperature anomaly and salinity driven overturning cell at the northern boundary.
in Figure 5.3 in contrast to the $\Delta S = 3.0$ case, the water of highest density for $\Delta S = 5.0$ is now completely separated from the surface, and below a depth that increases with latitude stable stratification is found. This is probably the cause of the sloping meridional transport contours of Figure 5.4a. The densest water of the upper 500 meters is found at latitudes between 47° and 59°N. When considering the surface density gradient, its direction can no longer merely be from south to north, but from north to south in the northern part of the basin. The nature of this density gradient explain the two opposite directed transport cells. The freshwater flux is now of such an extent that the water of highest density now longer can be found at the surface, but prevails in the abyss. The only way the density of the deep water can be altered are through diffusion and heat conduction, which are slow processes. Thus, letting the simulation run for a longer period of time should result in a deepening of at least the positive circulation cell.

As a consequence of the reduced transport of warm water towards high latitudes, the surface layer is cooled here, while the rest of the ocean domain experiences substantial heating due to the shut down in production of cold deep water (Figure 5.4c). Actually this choice of $\Delta S$, results in the greatest negative temperature anomaly of all experiments conducted. Another interesting feature of this case is a second positive overturning cell that has appeared at a depth below 3000 meters, confined to the southern part of the basin. This cell has a maximum transport of 2.6 Sv, and is a returning feature of all runs where $\Delta S$ in (4.3) exceeds 4.

Increasing the freshwater flux further causes the cell of negative overturning to grow both horizontally and vertically. The deep water remains the densest water of the whole ocean domain, and thus causes both the negative overturning cell as well as the remaining positive cell to be confined to the upper 1000 meters. As a consequence, the positive cell has to retreat southwards. Its vertical extent decreases with increasing freshwater flux as well. When a salinity forcing of $\Delta S = 8$ is applied the meridional streamfunction yield two well-defined separate circulation cells of opposite signs, as shown in Figure 5.5a. Both cells are associated with only weak transport, the positive cell being the strongest of the two. The two separate circulation cells indicate that the surface flow north of 40°N is directed south, while the corresponding flow south of this latitude is directed north. In the region between 36°N and 40°N the surface waters of the southern and northern part of the basin converges and sinks. This is in agreement with a density gradient of the nature described above. When meeting water masses of equal potential density the water spreads out horizontally, and thus the two circuits are closed. In the abyss at the southern boundary the positive overturning cell mentioned above prevails. It is in this cell we find the strongest transport in this run, with a maximum of 2.6 Sv.

The intrusion of warm water below a surface layer of cold low-salinity water is now well developed, and the warming of the ocean in its entirety is obvious. However, a cap of cold water remains in the upper 500 meters of the northernmost part of the basin (Figure 5.5b). As can be seen in Figure 5.5c the upper 100 meters of this cap is colder than the corresponding region in the verification run. When comparing the temperature anomaly plot with the one of the case where $\Delta S = 5$, we find that the negative temperature anomaly now have a greater horizontal extension. However, the region is approximately only half as deep. This is due to the heating of the remaining ocean domain. The negative overturning cell transports water of a higher temperature towards the surface at the northern boundary and thus effects the temperature here. The surface flow directed from north to
Figure 5.5: The three panels show results from an experiment in which $\Delta S = 8.0$ in equation 4.3. Panel a) shows the MOC (Contour interval 2.0 Sv, but 1.0 Sv and 0.5 Sv contours (both positive and negative) have been included as well), panel b) shows the zonal mean temperature (Contour interval 1.0°C), and panel c) show the temperature anomaly in the upper 500 meters (Contour interval is 0.2°C for negative anomalies and 1°C for positive anomalies). In panel a) cells of positive overturning are indicated by red shading while cells of negative overturning are shaded blue. Likewise, negative anomalies in panel c) are indicated by blue shading, while positive anomalies are shaded red. Note the two overturning cells of opposite signs and the negative temperature anomaly at the northern boundary.
south causes the negative temperature anomaly to spread out horizontally.

Adding more freshwater allows the negative overturning cell to continue growing horizontally, until the positive cell vanishes completely, and the meridional streamfunction is dominated by a single salinity driven overturning cell. This is first seen when $\Delta S = 10$ in (4.3) (Figure 5.6a). As the water sinking at the southern boundary has a density considerably lower than the sinking water of the verification run, this cell remains confined to the upper 1000 meters. Allowing the model to run for a longer period of time would probably deepen the circulation cell. However, the vertical growth of this cell will be very slow, as it is driven primarily by diffusion and heat conduction. The overturning strength is still weak, the negative cell has a maximum transport value of only 2.4 Sv.

The temperature field (Fig. 5.6b) now show a vigorous tongue of warm water intruding under a colder surface layer. At 1000 meters depth the temperature at northern boundary is as high as 16°C. This is not surprising when considering the flow pattern. Water of a temperature of about 27°C sinks at the southern boundary and spreads out horizontally. Thus we have a flow of warm water directed from south to north at intermediate depths. This leads to a substantial heating of the whole ocean domain. The temperature anomaly still give evidence of a cooling of the surface layer in the northern part of the basin. This negative anomaly is however confined to the upper 50 meters approximately, and its magnitude is $-1.2^\circ$C, which is 0.2°C warmer than for the $\Delta S = 8$ case. This proves that as the strength of the salinity driven circulation increases, the cooling of the surface waters due to the shut-down of the positive overturning will be neutralized. This finding is supported by an experiment in which the freshwater flux of $\Delta S = 15$ was applied, and additionally, to strengthen the circulation, the maximum salinity value was increased from 36 to 40. The temperature anomaly plot (Figure 5.7) for this case actually show a heating of 0.5°C at the northern boundary. A negative anomaly remains, however, with a maximum cooling of $-0.9^\circ$C. The region of cooling is very shallow, but stretches from 24°N to 60°N. The strength of the overturning circulation has increased substantially from the run where $\Delta S = 10$, the negative overturning cell now has a strength of $-24.8$ Sv. Its vertical extent is still restricted, but a deepening has occurred, it now covers the upper 1500 meters of the ocean basin.

The streamfunction in depth-density coordinates for all runs presented above, as well as the verification run, are presented in Figure 5.8. For the verification run this stream function show a strong thermohaline driven cell reaching all the way from surface to bottom, and involving only water with potential density larger than $\sigma = 27.3$. A positive cell indicates mechanically driven circulation of water with density in the range $\sigma = 27.0 - 27.6$ at depths between 500 and 1500 meters. A second thermohaline driven cell involving water of densities between $\sigma = 26.0 - 27.0$ is found in the upper 500 meters. The water in this cell is the lightest water of this run. Additionally, a few weak cells, of both anti- and clockwise circulation can be seen. The results is similar to the Nycander verification result presented in Figure 4.9. It should be noted that the resolution of the basin at hand is quite coarse and this affects the results. For the uniform temperature experiment the depth-density streamfunction consist of solely two cells. The dominating cell is thermohaline driven, and reaches from surface to bottom. The water involved in the circulation represented by this cell is heavier than $\sigma = 28.0$. The second cell appear to be mechanically driven, and have no contact with the surface. The water of lowest density in this run is included in this circulation cell. The streamfunction from $\Delta S = 3$ show a considerable weakening from the
Figure 5.6: The three panels show results from an experiment in which $\Delta S = 10.0$ in equation 4.3. Panel a) shows the MOC (Contour interval 2.0 Sv, but 1.0 Sv and 0.5 Sv contours (both positive and negative) have been included as well), panel b) shows the zonal mean temperature (Contour interval $1.0^\circ$C), and panel c) show the temperature anomaly in the upper 500 meters (Contour interval is $0.2^\circ$C for negative anomalies and $1^\circ$C for positive anomalies). In panel a) cells of positive overturning are indicated by red shading while cells of negative overturning are shaded blue. Likewise, negative anomalies in panel c) are indicated by blue shading and positive anomalies by red shading. Note the salinity driven overturning cell stretching from north to south and the intrusion of warm water under the cold layer of low salinity water.
verification run. It is still dominated by a thermohaline circulation cell covering the entire vertical range. Additionally, two mechanically forced cells are visible, neither of them are in contact with the surface. As for the meridional streamfunction there are substantial changes when the freshwater flux is increased to $\Delta S = 5$ (Fig. 5.8e). The dominating thermohaline cell of the previously presented streamfunctions no longer reaches the bottom. Also, the densest water of the ocean domain is not included by this cell. Two mechanically driven circulation cells have appeared, one at intermediate depths, stretching from 1000 - 2000 meters depth, involving water of a potential density between 25.4 - 25.7, and the second at a depth below 3000 meters, involving the densest water (25.6 - 26.0). The water of lowest density in this run seems to be involved in a mechanically driven circulation cell which occupies the upper 500 meters in a density range 23.75 - 24.5. When this density is reached the cell submerges under the dominating thermohaline driven cell.

As the freshwater flux is further increased the main thermohaline cell is weakened both in respect to strength and vertical extent. The densest water is no longer involved in this cell, which supports the results of the meridional streamfunction. It seems that as the freshwater flux increases the density of the surface waters is lower than the density of the water present at the beginning of the simulation (5°C and a salinity of 33). The densest water remains beneath the vertical level of 3000 meters, and is part of a mechanically driven cell. This probably correspond to the cell of positive overturning found at the same depths in the southern part of the ocean domain in the meridional streamfunction plots. The lightest water is also still involved in a mechanically driven circulation cell in the near-surface region.
5.1 Constant depth case

a) $\Delta S = 1.5$

b) $\Delta S = 1.5, \Delta T = 0$

c) $\Delta S = 3.0$

d) $\Delta S = 5.0$

e) $\Delta S = 8.0$

f) $\Delta S = 10.0$

Figure 5.8: The development of the streamfunction in depth-density coordinates for the cases discussed in this section. Contour interval is 5.0 Sv in Fig. 5.8a, and 3.0 Sv in Fig. 5.8b,c, 1.0 Sv in Fig.5.8d,e and 0.5Sv in Fig. 5.8f.
Figure 5.9: The development of the meridional overturning as $\Delta S$ is increased. Contour interval is 2.0 Sv in all figures. Additionally the 1.0 Sv and 0.5 Sv (both positive and negative) have been included.
5.1 Constant depth case

a) $\Delta S = 7.0$

b) $\Delta S = 8.0$

c) $\Delta S = 9.0$

d) $\Delta S = 10.0$

e) $\Delta S = 15.0$

f) $\Delta S = 15.0$, $S_{MAX} = 40$

Figure 5.10: The development of the meridional overturning as $\Delta S$ is increased. Contour interval is 2.0 Sv in all figures. Additionally the 1.0 Sv and 0.5 Sv (both positive and negative) have been included.
### Table 5.1: The meridional transport and temperature anomaly extremes for the different runs.

<table>
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<th>∆S</th>
<th>Meridional transport (Sv)</th>
<th>Temperature anomaly (°C)</th>
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<td>Minimum</td>
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<td>−1.1</td>
</tr>
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<td>2.8</td>
<td>−24.8</td>
</tr>
</tbody>
</table>

**5.2 Bottom topography case**

To investigate the effect of bottom topography results from experiments performed for the basin described in Nycander et al. (2007) are presented here. In order to simulate a freshwater flux into this basin salinity had to be added as a tracer. A boundary condition similar to (4.4) was applied, with the same restoring time scale. (5.1) yields a salinity forcing of the shape displayed in Figure 4.1.

\[
S = 36 + \frac{\Delta S \left[ \cos\left(\frac{\pi y}{L}\right) - 1 \right]}{2},
\]  

(5.1)

An experiment where \( \Delta S = 1.5 \) in the equation above, was conducted for both a horizontal resolution of 100 km × 100 km and 200 km × 200 km. The configuration is identical to...
Figure 5.12: The six panels show results from an experiment in which $\Delta S = 1.5$ in (5.1). The left-hand panels display results from a run where the horizontal resolution is $100 \text{ km} \times 100 \text{ km}$, while the right-hand panels show results from a run with a horizontal resolution of $200 \text{ km} \times 200 \text{ km}$. Panel a) and d) show the MOC (contour interval 0.5 Sv), panel b) and e) show the zonal mean temperature (contour interval 1.0°C), and panel c) and f) show the temperature anomaly (contour interval is 0.2°C). In panel c) only the upper 100 meters are included, while the upper 200 meters are included in panel f). Note the negative temperature anomaly at the northern boundary.
Figure 5.13: Zonal mean salinity a) The high resolution $\Delta S = 1.5$ run and b) The low resolution $\Delta S = 1.5$ run. Contour interval is 0.2. As the salinity field essentially is confined to the surface layer, only the upper 1500 meters of the basin is included.

the verification run described in Section 4.2 with the exception of the equation of state in the low resolution run, which had to be nonlinear in order to ensure numerical stability. Additionally, salinity forcing was added at the surface. The high resolution run is presented first. The meridional overturning streamfunction for this run is shown in Figure 5.12a. The most striking feature is that the deep cell of Figure 4.8a is almost completely gone. Only a small cell at the bottom in the southern end of the basin remains. At first sight it is difficult to spot any significant changes in the upper circulation cell. However, some differences are visible. The 1 Sv-contour stretches further north in the present simulation, and the 0.5 Sv-contour sinks further south than earlier. As in the verification run this contour continues north below the surface, but it does not reach as far north in the present experiment, supporting the theory on weaker overturning when freshwater is added in the northern region. At intermediate depths, however, this contour stretches further north. Also, a tiny cell of positive anti-clockwise circulation, barely visible in the verification run, can be seen at the northern boundary. The zonal mean potential temperature (Figure 5.12b) shows evidence of a slight warming, 3$^\circ$C-contour extends down to approximately 2000 meters, compared to a maximum depth of about 700 meters in the verification. But the most interesting feature of this plot is the northward slope of the contours in the first model layer. This occurs at the same meridional position as the sinking of the 0.5 Sv-contour mentioned above, and indicates cold, low salinity waters forcing the warm water flowing north to sink, an continue its way north below this layer. The temperature anomaly plot (Figure 5.12c) shows a negative temperature anomaly in the upper 30 meters at the northern boundary. A discontinuity is found right off the shelf, this can be explained by examining the zonal mean temperature at the same location, were water warmer than its surroundings is found. At the most the temperature is $-1.48^\circ$C colder here than in the verification run. The ocean basin as a whole experiences a slight heating.

This is the only experiment with an added freshwater flux conducted with the same resolution as applied in Nycander et al. (2007). These experiments qualitatively match those of higher resolution and thus is sufficient to investigate the effect of topography. In addition
Figure 5.14: The three panels show results from an experiment in which $\Delta S = 8.0$ in equation 5.1. Panel a) shows the MOC (Contour interval 0.5 Sv), panel b) shows the zonal mean temperature (Contour interval 1.0°C), and panel c) show the temperature anomaly in the upper 200 meters (Contour interval is 0.2°C. Note the negative temperature anomaly at the northern boundary.
these runs are more effective on the computer. The results from the same experiment, but coarser resolution is displayed in Figure 5.12d,e,f. The strength of the intermediate cell has increased, whereas the strength of the upper cell has decreased. Whether the 1.0 Sv or the 0.5 Sv-contour reach as far north as in the verification run. The latter covered the entire meridional length of the basin in the verification run, but now it loses contact with the surface approximately 2000 km from the northern boundary. Additionally, a deep cell stretching from a depth of 1500 meters and all the way to the bottom has appeared at the northern boundary. As in the high resolution run, the zonal mean temperature plot (Figure 5.12e) show evidence of the ocean basin being heated. The coldest water is now confined to the northern boundary. The temperature anomaly supports this finding, but cooling can be found at the northern boundary, reaching as deep as 150 meters which is the shelf depth. The maximum deviation is $-0.62^\circ$C, which means that the predicted cooling is greater in the high resolution run. The mean heating of the ocean as a whole is however greater in the low resolution run.

Increasing the salinity forcing further yield only minor changes on both the streamfunction and temperature fields. Figure 5.14 shows the meridional overturning streamfunction, zonal mean temperature field and the temperature anomaly in the upper 200 meters of the basin for an experiment where $\Delta S = 8.0$ in (5.1). The changes in the overturning circulation (Figure 5.14a) are insignificant. The temperature plot show the same trends as for the $\Delta S = 1.5$ case, the slope of the contours at the first model layer are steeper now, and can be seen for higher temperatures than before. The area of negative temperature anomaly is now more horizontally aligned than before, it is now confined to the upper 20 meters. This fact is in agreement with the results from the experiments in the basin of constant depth. The distribution of salinity in the basin provides an explanation for the lacking response to the added freshwater. Figure 5.15 shows the salinity field for the case $\Delta S = 8$. The freshwater flux fails in spreading in the vertical and remains in the surface layer. Only the upper 50 meters can be said to be stratified with respect to salinity. The salinity fields for both runs where $\Delta S = 1.5$ are shown in Figure 5.13. They both indicate
that the salinity stratification is more or less confined to the shelf depth. However, the response for this freshwater flux is stronger than for the $\Delta S = 8$ experiment.

Figure 5.16 shows the streamfunction in depth-density coordinates for all three experiments presented above. For the high resolution $\Delta S = 1.5$ run the densest water does not reach the same depths as in the corresponding verification run, and a small anti-clockwise cell centered at a depth of 2000 meters has appeared (Figure 5.16a). Apart from this the agreement of the streamfunctions for the two runs is very high. As the meridional streamfunction show only minor changes between these runs, no major changes should be expected either. The differences between the two corresponding low resolution are greater. Here, it should be reminded that the equation of state used in the verification run is not the same as in the freshwater flux experiment and could be the cause of some changes. The latter experiment yield a more chaotic streamfunction and a new feature is that the densest water do not appear at the surface. As for the high resolution run a anti-clockwise circulation cell not found in the verification run has appeared in the abyss. The $\Delta S = 8$ run yield a smoother streamfunction than the previous case. The positive cell that appeared at depth in the latter case, has vanished here. The densest water is not present at the surface in this case either, and the circulation does not reach the bottom anymore. A feature not seen in any of the depth-density streamfunction plots discussed before is the long “tail” of low density water confined to the top level. This is probably due to the low salinity water on the northern boundary, which fails to spread out vertically.
Figure 5.16: The streamfunction in depth-density coordinates. Contour interval is 1 Sv. Red shading indicates anti-clockwise circulation while blue shading indicates clockwise circulation. a) High resolution $\Delta S = 1.5$ run. b) Low resolution $\Delta S = 1.5$ run. c) Low resolution $\Delta S = 8.0$ run.
Chapter 6

Discussion

The North Atlantic is an important region for the thermohaline circulation of the world ocean as it is one of few locations where deep water is formed. The surface water must attain sufficient high density for the underlying water column to become neutrally stratified and thereby allowing the water to sink from the surface and into the abyss. The density at the surface is altered through cooling and when sea ice forms, by the rejection of brine from the ice. As a consequence of global warming the amount of freshwater added to this region is expected to increase and thus cause the surface water here to become less saline. This could be crucial for the deep water formation and it is debated whether a shutdown of the deep water formation would lead to a shutdown of, or even a reversal of the meridional overturning circulation.

The theory regarding the thermohaline circulation given in Chapter 2 based on simple one and two-dimensional models suggests that two regimes of flow can exist, one where the density field is dominated by temperature and one where salinity is the dominating effect on density. Our experiments with a three-dimensional ocean of constant depth suggest a coexistence of these two flow regimes, with temperature driven overturning in the southern part of the basin and salinity driven in the northern part. The overturning strength predicted is however considerably weaker than in the reference state, and the depth of the overturning cells is quite shallow, due to the initially dense water of the configuration. This is consistent with what could be expected if the density of the sinking water was to decrease, as the density of the deep water is correspondingly high.

As the freshwater flux at high latitudes is increased, our experiments are consistent in predicting a weakening of the temperature dominated overturning. Associated with this weakening is a decrease of surface layer temperatures at the northern boundary, and a heating of the remaining ocean domain due to reduced production of cold deep water. For our flat bottom ocean the overturning remains positive even when the salinity at the northern boundary is decreased from 34.5 to 32.0. The strength of the overturning is now weaker than the reference state by 56%, and is associated with substantial cooling of the surface layers. When considering surface salinity in the North Pacific, values in the the range 30.0 - 32.0 can be found in near-coastal regions (Pickard and Emery, 1990). There is no evidence of deep sinking occurring in this ocean as the low salinities poses a restriction on the density. If the salinity of the North Atlantic water was decreased to such an extent, we could expect severe impact on the deep water formation taking place here.
The results from the experiment where $\Delta S = 5.0$ in (4.3) resembles the current situation in the Arctic Sea. Here, warm Atlantic water of high salinities enters at depths from 200 to 900 meters (Pickard and Emery, 1990), while low-salinity water with temperatures below zero, leaves the region as surface currents. The flow pattern found in our experiments is a result of surface density gradients. As the densest surface water no longer is found at the northern boundary, the density gradient is directed from north to south at this boundary. At the equator, however, the direction of the density gradient remains unaltered. This means that a southward surface flow is enforced at high latitudes, and an opposite directed surface flow is induced at low latitudes (Figure 6.1). The zone in which these flows converge is characterized by downwelling. The heat transported to the northern boundary is reduced due to these circulation changes. This is reflected in the negative temperature anomalies found at the northern boundary in all experiments with an increased freshwater forcing.

Further increasing the freshwater flux causes the location of densest surface waters to move south and thus, the convergence zone moves southward as well. As the southward surface flow of cold, low-salinity water reaches further south, the negative temperature anomaly spreads out horizontally too. Simultaneously, the shutdown in production of cold deep water causes the mean temperature of the ocean to increase. This means that the temperature of the water upwelling at the northern boundary increases, and thus the vertical extension of the negative temperature anomaly is diminished. This has an impact on the magnitude of the anomaly as well, the maximum negative temperature anomaly increases with increasing freshwater flux until this flux exceeds $\Delta S = 5.0$ in (4.3). After this level is reached the negative temperature anomaly decreases with increasing $\Delta S$. If the freshwater flux was further increased it is likely that eventually the temperature anomaly would be positive for the entire ocean domain.

Adding topography causes the overturning cell to become shallower and a reduction of the overturning strength. The presence of topography introduces new dynamics to our basin, as the flow tends to be parallel to the $f/H$-contours when topography is included. The
Figure 6.2: Horizontal velocity superimposed on a) the horizontal temperature distribution and b) the $f/H$-contours for the low resolution flat bottom experiment without salinity forcing. Both velocity and temperature are taken at a depth of 20 meters. Note the eastward jet at the western boundary, approximately 2200 km from the southern end, and how the flow affects the temperature here. It is also of interest to see how the flow chiefly is parallel to the $f/H$-contours. The reference vector (blue arrow at the top) represents a velocity of 25 cm/s.
topography also impose a restriction upon the deep sinking. The dense water formed at
the surface can only sink to the depth of the water column in which it was formed, and this
depth is shallow at the boundaries where dense water is formed. In his numerical study
of the circulation in an ocean basin similar to our bottom topography basin (although in
this study no topography was included at the equator) Winton (1997) argues that a larger
portion of the thermohaline circulation takes place in the horizontal plane rather than in
the meridional plane when topography is included. This seems to be consistent with our
results, as the surface velocities in the topography experiments are greater than the velo-
cities in the constant depth experiments by an order of magnitude (Figure 6.2). Another
reason for the weaker overturning in the bottom topography experiments is the intense
mixing at the boundaries in the constant depth experiments. This parameterization fa-
vors mixing of the water properties at the northern and southern boundary. That is, the
lightest and densest waters are mixed to a greater extent than water of other properties.
Applying uniform mixing in this basin yield a weaker overturning as shown in Marotzke
(1997). An experiment conducted with ROMS applying uniform mixing in the constant
depth configuration yields similar results (not shown). Adding a freshwater flux to the
bottom topography basin yield a weakening of the meridional overturning, but this weak-
ening is small compared to the results from the constant depth basin. The zonal mean
salinity profiles provide an explanation for the resistance of this basin, as they show strat-
ification due to salinity in only a shallow surface layer. The experiments with $\Delta S = 1.5$
in (5.1) indicate that the shelf depth pose a restriction on the vertical salinity distribution.
Conducting the constant temperature experiment described in Section 5.1 for the bottom
topography configuration results in no meridional overturning at all (not shown), indic-
ating that a salinity driven overturning circulation can not exist in this case. The weak
response of this basin to an increased freshwater flux should be further investigated.

As all experiments presented in this thesis apply solely thermohaline forcing, we expect the
circulation to appear as thermohaline driven in the depth-density streamfunction. This is
overall true, although mechanically driven cells appear in all plots, the dominating cells
are thermohaline driven. The shallower meridional overturning for the freshwater flux
experiments manifests itself also in the depth-density streamfunction, as the dominating
thermohaline driven cells no longer cover the entire ocean depth. An interesting feature is
the temperature dominated overturning cell in the abyss, found in all experiments where
$\Delta S \geq 4.0$ (see Figure 5.9;5.10). The depth-density streamfunction diagnoses this cell as
mechanically driven, meaning that the water sinking in this circulation is lighter than the
rising water. The water at this location is however stable stratified with only minor density
variations. Thus the resolution of density in the numerical routines used to calculate $\psi^*$
may have great influence on the results, and the reliability of the depth-density streamfunction
should be questioned here. It should also be taken under consideration that the resolution
of the constant depth configuration is coarse, both horizontally and vertically. These
experiments should thus be repeated with better resolution in order to rule out spatial
resolution as a source of error.

The changes on the ocean circulation presented in Chapter 5 and discussed above could
have a vigorous impact on the climate. The negative temperature anomalies together
with the fresh surface water at the northern boundary provide conditions for increased
sea ice formation, and may allow the ice cover of the Arctic to extend further south.
This could cause a cooling of northern Europe. It should be noted that the negative
temperature anomalies decreases with depth and as the first model layer is positioned 17 meters below the surface in our constant depth configuration, even greater anomalies should be expected at the surface. Our freshwater flux is simulated through forcing the surface salinity toward the profiles given by (4.3) and (5.1), which is a crude approximation. The sea-level rise due to increased precipitation and melting of glaciers thus are lost in our simulations. The heating of the ocean domain as a whole would also contribute to a sea-level rise due to thermal expansion. ROMS is, however, a Boussinesq model and thus volume is conserved in our simulations. ROMS is therefore incapable to predict sea-level rise due to thermal expansion. In Manabe and Stouffer (1994) the response of a coupled ocean-atmosphere model to an increase of CO$_2$ in the atmosphere, the predicted sea-level rise due to thermal expansion was 1-2 cm. In some of our experiments the heating of the ocean is severe, and the expected sea level rise due to thermal expansion should thus be expected to exceed their findings. Another flaw with our freshwater flux is that it is applied at every time-step throughout the simulation. Global warming, however, would induce a freshwater flux in the form of a pulse, on a decadal time-scale. This freshwater flux would then act upon an already existing overturning circulation and weaken it.

Wind forcing is omitted in this thesis. Wind stress is an important driving force for surface currents such as the Gulf Stream, which is a major part of the meridional overturning circulation in the North Atlantic. Including wind forcing would lead to an intensification of the already intense western boundary current. This would enhance the positive overturning, and should thus make the overturning circulation more resistant to changes induced by an increased freshwater flux. Furthermore, the presence of topography would also impose a restriction on the changes on the meridional overturning circulation induced by an increased freshwater flux. Global warming would also result in increased evaporation at low latitudes and thus the import of salinity to high latitudes should increase as well, and thereby limiting the effect of the freshwater flux.

In order for the meridional overturning circulation to shutdown the salinity of the surface waters at high latitudes must be considerably lowered. Such changes are unrealistic. However, slight changes in surface salinity lead to a weakening of the circulation, but it is not likely that these changes would be able to override or even counteract the global warming induced temperature changes. The freshwater flux required to induce a salinity dominated overturning circulation is unrealistic, and thus a circulation of this nature can be ruled out.
Chapter 7
Conclusions

In this study the impact of an increased freshwater flux on the meridional overturning circulation in an ideal version of the North Atlantic have been examined. The freshwater flux was introduced to the ocean by forcing the surface salinity towards the profiles given by (4.3) and (5.1) and visualized in Figure 4.1. The main focus has been on the meridional overturning circulation and the temperature anomalies caused by the changes on the circulation. The streamfunction in depth-density coordinates introduced by Nycander et al. (2007) have been used in order to analyze the energetics of the circulation at hand. ROMS, a modified terrain following vertical coordinate model, was used to perform the simulations this thesis is based on. Numerical studies of ocean climate have in the past been carried out mainly with the use of geopotential coordinate models, and thus it is also of interest to see how well the results from a terrain following vertical coordinate ocean model correspond to earlier simulations.

In Chapter 4 the results from Marotzke (1997) and Nycander et al. (2007) were reproduced by the use of ROMS. This verifies that ROMS may indeed be used to investigate the meridional overturning circulation. The main difference of these configurations is the presence of bottom topography in the latter. This is essential, as in the case of constant depth ROMS will in fact act as a geopotential model. Overall, the agreement of results from the ROMS experiments and the results given in Marotzke (1997) and Nycander et al. (2007) is good. In the latter case, however, there are some differences, but this should be expected as some changes on the configuration had to be made in order to achieve numerical stability. Thus, we conclude that ROMS indeed is an adequate model for studies of the meridional overturning circulation.

Results from the experiments with an increased freshwater flux are given in Chapter 5 and discussed in Chapter 6. The results from the constant depth configuration show that a complete shutdown, and even reversal, of the meridional overturning circulation is possible. The amounts of freshwater that must be supplied to the ocean at high latitudes are, however, substantial and unrealistic. Thus, a salinity dominated overturning circulation as a result of global warming is highly unlikely. However, increasing the freshwater flux only slightly leads to a weakening of the circulation, and associated with this weakening is a decrease in surface layer temperatures at the northern boundary. The reduced production of cold deep water will cause the remaining ocean to be heated.
The meridional overturning circulation in the basin with bottom topography seems to be more resistant to changes in the freshwater flux. The results are, however, consistent with the constant depth experiments in predicting a weakening of the circulation, and a decrease of surface layer temperatures in the northern part of the basin.

Applying the streamfunction in depth-density coordinates as a tool to analyze the energetics driving the circulation yield overall good results. Thermohaline driven cells are dominating the streamfunction in all experiments, but mechanically driven cells can be found in all cases. This may be due to resolution, both spatial in the model and the intervals we choose to split potential density into. Parameterization of subgrid processes in the model may also affect the results to some degree.

Based on these findings we conclude:

- An increased freshwater flux will indeed have a weakening effect on the meridional overturning circulation.
- As a result of this weakening the surface layer at high latitudes will experience a cooling, while the remaining ocean will be heated. It is however unlikely that this cooling will be severe enough to induce a cooling of the North European climate under global warming. The heating of the ocean as a whole may, however, lead to sea level rise due to thermal expansion.
- The presence of bottom topography imposes a restriction on these changes.
- If the ocean was supplied with sufficient amounts of freshwater a complete shut-down and reversal of the thermohaline circulation is a possibility. The amounts of freshwater needed are, however, so large that this scenario may be excluded.
- The use of the streamfunction in depth-density coordinates yields good results, but effort should be made to improve its reliability.

As our experiments have been conducted for an ideal ocean with forcing of a simple nature, the uncertainties are large. In order to minimize these uncertainties, experiments with a more realistic ocean basin applying more realistic forcing should be conducted. In particular, wind forcing should be included, as this is an important driving force for the surface currents. In particular, experiments where the resolution of the constant depth configuration is increased, both horizontally and vertically, should be performed in order verify the existence of the deep cell in the abyss at the southern boundary and the mechanisms driving this cell.
Bibliography


