Lecture 24. Thermohaline Circulation: multiple states and catastrophe  

1. Thermohaline circulation under mixed boundary conditions  
   A. F. Bryan technique:  
   First, run the model to quasi-equilibrium state under the relaxation conditions for both temperature and salinity  
   \[
   \frac{dT}{dt} = \Gamma(T^* - T), \quad \frac{dS}{dt} = \Gamma(S^* - S)
   \]
   Second, when the model reaches a quasi-equilibrium, the equivalent salt flux at the upper surface is diagnosed,  
   \[S_f = \Gamma(S^* - S)\]
   Third, the model is restarted from the quasi-equilibrium state and run under the mixed boundary conditions.  
   \[
   \frac{dT}{dt} = \Gamma(T^* - T), \quad \frac{dS}{dt} = S_f,
   \]
   i.e., a relaxation condition for temperature and a flux condition for salinity. Note that the salt flux is now considered as fixed. Using this technique, it is much easier to find multiple solutions in OGCM.

   In F. Bryan's experiments, he had to add some salinity perturbation on the poles to get the catastrophe started. Recently, most numerical model experiments do not need such initial impact, and the model's solutions automatically drift away and reach to the other solutions.

B. Potential problems:  
1) Relaxation for salinity is unphysical, so \(\Gamma\) is artificial anyway. Tzipermann et al. (1994, J.P.O.) argued that a small value of \(\Gamma\) should be used; however, this does not solve the problem.  
2) The virtual salt flux used in the model is unphysical. In fact, there is no salt flux cross the air-sea interface; thus, it is much better to use the natural boundary condition for salinity.  
3) The salt flux diagnosed in the model has no physical meaning when the model is still in the process of reaching a quasi-equilibrium.  
4) As indicated by many numerical experiments, the solutions are very sensitive to the exact shape of the virtual salt flux profile, so a more accurate boundary condition should apply.  
5) Most numerical experiments based on such approach give rise to virtual precipitation over the Gulf Stream outflow region where evaporation should prevail due to the strong heat loss to the atmosphere.

C. Halocline catastrophe by F. Bryan (1986)  
   (1) What is a halocline?  
   Halocline is defined as a layer where the vertical salinity gradient is local maximal. Halocline exists in the world oceans, mostly as the interface between a layer of low salinity water on the upper surface and the relatively salty water below. Outstanding permanent halocline exists in the subpolar gyre in the North Pacific Ocean and the Arctic Ocean (Fig. 1). Formation of the halocline in these locations is closely related to excess precipitation or local converge of relatively fresh water in the upper ocean.

   In addition, non-permanent halocline can developed due to different dynamical processes:
Diurnal heating and rain events may substantially influence the turbulent motions in the upper ocean and lead to the formation of a halocline in the upper ocean (Soloviev and Lukas, 1996). Freshwater cap may appear in the subpolar Northern North Atlantic, which may be induced by either import from freshwater from melting ice or excessive precipitation over the subpolar basin.

Rooth's speculation of the stable asymmetric pole-pole circulation was confirmed by F. Bryan's (1986) numerical experiments based on the GFDL primitive equation model. At that time it was not an easy job to find the pole-pole circulation. Bryan started with a two-hemisphere model ocean with symmetric forcing, with relaxation condition for both the temperature and salinity. A circulation symmetric with respect to the equator was established.

In order to find the asymmetric circulation he tried many things to trigger the collapse of the symmetric circulation. For example, he tried to impose a large positive salinity anomaly at high latitudes, but the symmetric circulation seems to be insensitive to such perturbations. However, he found that if a relatively large negative salinity anomaly was imposed at the northern limit of the model ocean, a rapid response of the meridional circulation took place. The collapse of the circulation is apparently caused by the interruption of deepwater formation at high northern latitudes due to the existence of the new halocline. Since this kind of abrupt change is induced by halocline, it is now called the **halocline catastrophe**.

In the process of searching such a catastrophic change in the thermohaline circulation, F. Bryan developed a technique, which has been widely used since then. This technique is as following. When the system reaches a quasi-equilibrium, the equivalent salt flux required for maintaining the salt distribution is diagnosed and this salt flux is used as the forcing for the salinity at the second stage of the numerical experiments. Since the relaxation condition still applies to the temperature field, the new boundary condition has been called the mixed boundary conditions.

Bryan found that the steady solution under the relaxation condition for the salinity is not stable under the new mixed boundary conditions and a polar halocline catastrophe set in. As the result, the thermohaline circulation collapses and an asymmetric pole-pole circulation is established gradually, Fig. 2. The entire solution changes accordingly. Most importantly, the poleward heat flux alters, Fig. 3.
D. Implications of F. Bryan’s numerical experiment

The thermohaline catastrophe has been observed in many numerical experiments based on OGCM during the past few years. The renewed interest on the thermohaline circulation is due to its close link with the climate. Ocean covers 70% of the Earth's surface. Water has a much high heat capacity compared with the air; thus, the climate on a planet without the oceans would be extremely different from a planet with oceans. According to observations, heat flux in the oceans contributes about 50% of the total poleward heat flux.

However, as shown in the numerical experiments by F. Bryan, this poleward heat flux can be changed substantially if the thermohaline circulation in the oceans collapses. The oceans are so big, it is very hard to image that the oceanic circulations can ever change. However, according to paleoclimate records, oceanic circulation did change many times in the past.
The most remarkable example is the so-called Younger Dryas period, which is about 11,000 years ago. There are strong evidences that suggest that during the last glaciations the thermohaline circulation was interrupted due to the extensive ice-coverage over the northern North Atlantic. At the end of the last glaciations the thermohaline circulation gradually returned to a situation rather similar to the present-day situation. However, during a very short period, the so-called Younger Dryas period, the thermohaline circulation in the North Atlantic was interrupted again. The interruption of the thermohaline circulation and the sudden cooling of the European and North American Continents have been confirmed by many paleoclimate records.

It has been speculated that the large amount of fresh water from the ice melting flooded the subpolar basin of the North Atlantic. This cold and fresh water cap prevented the evaporation and cut down the heat flux from the ocean to the atmosphere. As the result, deepwater formation was interrupted, so was the thermohaline circulation. There have been many studies on the thermohaline circulation and halocline catastrophe.

At present, numerical models of the thermohaline circulation in the oceans cannot reproduce some of the basic structure of the circulation. For example, the thermocline reproduced in most numerical models is much too thick. The poleward heat flux in these models cannot match the observations, the strength of the thermohaline circulation is different from observation. As one example for these problems, let us examine what happened during the last glaciations. Fig. 4 shows a schematic picture of the thermohaline circulation reconstructed from $\delta^{13}C$ profile in the Atlantic Ocean during the last glacial maximum. It is to notice that there was very strong intermediate water formation that dominated the upper 2 or 3 kilometers. However, such kind of thermohaline circulation has not been reproduced by numerical models. Apparently, much work has to be done before the numerical models can simulate the oceanic circulation properly.

![Fig. 4. Reconstruction of the $\delta^{13}C$ profile in the eastern Atlantic Ocean during the last glacial maximum, 18,000 years ago. The $\delta^{13}C$ values of the total dissolved $CO_2$ have been estimated from the $\delta^{13}C$ values of benthic foraminifera genus Cibicides in 34 sediment cores. Surface $\delta^{13}C$ have been derived from those of N. Pachyderma using the correction factor of Labeyrie and Duplessy (1985) (Duplessy et al., 1988).]
2. The conveyor belt
A. The concept

The conveyor belt was first proposed by Broecker (1987) as an illustration in an article appeared in National History, Fig. 5. This conception is so appealing that the conveyor has become a hot topic, although not without hot debate! It was assumed that this conveyor is driven by sinking in the northern North Atlantic. Of course, the new energetic theory of the thermohaline circulation claims that such a conveyor must be maintained by mechanical energy from wind stress and tidal dissipation.

Fig. 5. The great ocean conveyor logo (Broecker, 1987) (Illustration by Joe Le Monnier, Natural History Magazine.)

There are many ways to estimate the sinking rate of the NADW. Broecker (1991, Oceanography) used the tracer balance as the follows. As radiocarbon decays by one percent in 82.7 years, the $^{14}C$ deficiency map for the Atlantic, Fig. 4.2 suggest the residence time of 83 - 250 years, so the average age is 180 years. The total volume of the deep Atlantic basin is $2500 \times 6.2 \times 10^5 = 1.55 \times 10^7 \text{ (m}^3\text{)}$. Thus, the ventilation flux is $V/T = 8.6 \times 10^{14} \text{ m}^2 \text{/ year} \approx 27 \text{ Sv}$.

The formation rate of the AABW is estimated to be 4 Sv, thus, the NADW formation rate is about 23 Sv. Note that the original source of the deep water formation rate in the Norwegian/Greenland Sea is only about 0.47 Sv; however, along the path of the deep western boundary current, there is much more water entrained into the deep western boundary current. The flow segment along the eastern coast of Greenland for density larger than $\sigma_b \geq 27.8$ has a total volume flux of 10.7 Sv (Dickson et al., 1990, Nature); when it reaches the southern tip of Greenland, the volume flux is increased to 13.3 Sv. Thus, Broecker's estimation is not inconsistent with the other observations.
The conveyor belt is an over simplified concept. Because of its simplicity many people use this term as a tool to describe the thermohaline circulation in the world oceans. However, the dynamic picture implied by this cartoon is so much over-simplified that it can be quite misleading.

One of the most critical problems hidden in this cartoon is the omission of the Antarctic Circumpolar Current, which plays the role of the global artery of the world ocean circulation. For example, a better picture of the global circulation was proposed by Bill Schmitz (1995). Accordingly, major part of the water mass transport has to go through the whole loop of ACC before it can return to the Northern North Atlantic and repeat the cycle again. During this process wind stress forcing in the Southern Ocean plays a vitally important role in regulating this global conveyer.
3. Multiple solutions for the global oceans
   A. Multiple solutions for the global ocean obtained from idealized model

   Marotzke and Willebrand (1991) carried out a series of numerical experiments for the global oceans based on the GFDL. The model consists of a highly idealized geometry, with two ocean basins of equal width and meridional length, Fig. 4.3. The model is first spun up under the relaxation condition for both the temperature and salinity. A global zonally mean SSS is used as the reference salinity. After the model reaches a quasi-equilibrium, the virtual salt flux is diagnosed and the model is restarted and run under the mixed boundary conditions. Since there are four sites where deep water can be formed; thus, theoretically, there can be $2^4 = 16$ possible patterns of deepwater formation. They found several modes of the global circulations, Fig. 9 and 10.
Fig. 9. A global model to simulate the global modes of the thermohaline circulation (Marotzke and Willebrand, 1991).

Fig. 10. Left: Equilibrium 1: northern sinking. a) Meridional streamfunction in Sv as the sum of both basins; b) zonally averaged salinity and c) northward heat flux in PW for both basin combined. Right: Equilibrium 4: southern sinking. a) Meridional streamfunction in Sv as the sum of both basins; b) zonally averaged salinity and c) northward heat flux in PW for both basin combined.
Fig. 11. Equilibrium 2: the conveyor belt. a) Atlantic and b) Pacific meridional streamfunction in Sv c) Atlantic and d) Pacific zonally averaged salinity; e) Northward heat flux in PW for Atlantic (dotted), Pacific (dashed), and the two basin combined (solid).

Note that they used the same global mean sea surface salinity independent of longitude, so their solutions for the Atlantic and Pacific are made symmetric, and their solutions cannot take into consideration of the vapor flux across Central America. Whether the mode with deepwater formation in the North Pacific can be realized in the real ocean is questionable.

B. Two equilibria for the Atlantic in the air-sea coupled system

Manabe and Stouffer (1988) carried out a numerical experiment based on the Princeton atmosphere-ocean coupled model. In their experiment, two equilibria were found. The first one labeled as EXP I represents the present-day climate, the second one labeled as EXP II is quite different. In fact, in the second experiment the North Atlantic is much colder and fresher (Fig. 5). The maximum temperature and salinity difference are $5^\circ C$ and 3.
C. Implication for climate change?

Paleo climate records indicate that the meridional circulation in the Atlantic basin has gone through cycles of On/Off, thus, the hottest issues in oceanic circulation and climate study is whether such an abrupt change in the North Atlantic circulation may happen down the road.

The main focus is the potential role of freshwater flux:

1. Due to global warming, hydrological cycle may be intensified, with more evaporation at low latitudes and more precipitation at high latitudes. The meridional salinity gradient can be enhanced, producing a meridional pressure difference against that due to thermal forcing. As a result, the meridional overturning cell associated with the NADW may be slowed down and even be interrupted. According to IPCC (2001) report, many of the state-of-art climate models predict that meridional overturning circulation in the Atlantic basin will be substantially reduced over the next 100 hundred years, Fig. 13.

2. Due to global climate change, within the next 30-50 years the Arctic Ocean may be ice free in summer time. Without the ice, the large amount of relatively fresh water in the Arctic Ocean may not be hold within the Arctic basin. If some water from this freshwater
pool floods the North Atlantic, it may cause the halocline catastrophe, similar to what has been simulated by F. Bryan and others.

Fig. 13. Change of overturning rate in the Atlantic predicted by 9 state-of-art climate models.

Note that these results obtained from models based on the assumption that diapycnal mixing coefficients are invariant under different climate conditions. From the view of the new energetic theory, wind stress, and to a mild degree the tidal dissipation, may be quite different under different climate conditions. Thus, results from such model experiments remain questionable. It is hoped that with the rapid progress along the line of unrevealing the mystery of mixing and oceanic circulation, we will be able to simulate and predict the oceanic circulation and climate more accurate in the near future.

4. Zonally-averaged model for the thermohaline circulation

A. The zonally-averaged model by Marotzke et al. (1988)

Consider an ocean of depth $H$ and width $D$, the zonally-averaged momentum equations are

\[ -f\tilde{v} = \frac{p_e - p_w}{\rho_0 D} + A \tilde{u}_{zz} \]  
\[ f\tilde{u} = \frac{1}{\rho_0} \tilde{y}_{yy} + A \tilde{v}_{zz} \]  

Eliminating $\tilde{u}$, we obtain an equation for $\tilde{v}$

\[ f^2 \tilde{v} + A^2 \tilde{v}_{zzz} = \frac{f}{\rho_0 D} (p_e - p_w) + \frac{A}{\rho_0} \tilde{v}_{yzz} \]  

(2)
In order to solve this equation we need a parameterization of the E-W pressure term, Marotzke et al. suggested that this term should be parameterized in term of the north-south pressure gradient; however, they did not carry out this parameterization. In fact, they took a short cut by assuming a different balance

\[ A \vec{v}_z = \frac{1}{\rho_0} \vec{p}_y \]  

(3)

Notice that this is equivalent to set \( f \equiv 0 \), so the Coriolis force drops out. In the zonally-averaged model, the continuity equation is

\[ \vec{v}_y + \vec{w}_z = 0 \]  

(4)

Thus, a streamfunction can be introduced

\[ \psi_y = \vec{w}, \psi_z = \vec{v} \]  

(5)

Assume the equation of state is a linear one

\[ \rho = \rho_0 (1 - \alpha T + \beta S) \]  

(6)

The hydrostatic relation is

\[ \vec{p}_z = -\rho g \]  

(7)

By differentiating (7) twice and using (1a), we obtain a single equation for \( \psi \)

\[ \psi_{zzzz} = \frac{g}{\rho_0 A} (-\alpha T_y + \beta S_y) \]  

(8)

There are two prognostic equations for the temperature and salinity

\[ T_i - \psi_z T_y + \psi_y T_z = \kappa T_{zz} + C(T) \]  

(9)

\[ S_i - \psi_z S_y + \psi_y S_z = \kappa S_{zz} + C(S) \]  

(10)

where \( C(T) \) and \( C(S) \) indicate the convective mixing due to overturning. The boundary conditions for this equation are

\[ \psi = 0, \psi_z = 0, \text{ at } z = 0 \]

\[ \psi = 0, \psi_z = 0, \text{ at } z = -H \]  

(11)

\[ \psi = 0, \text{ at } y = \pm L \]

which means no mass flux through the bottom and side wall, no-slip at the bottom and no wind stress on the upper surface.

The boundary conditions for temperature and salinity are no-flux condition at the solid boundary

\[ T_z = S_z = 0, \text{ at } z = -H; T_y = S_y = 0, \text{ at } y = \pm L \]  

(12)

On the upper surface, the Dirichlet boundary condition

\[ T = T_0 \left[ 1 + \cos \left( \frac{y \pi}{L} \right) \right], \quad S = S_0 \left[ 1 + \cos \left( \frac{y \pi}{L} \right) \right] \]  

(13)

or the mixed boundary condition

\[ T = T_0 \left[ 1 + \cos \left( \frac{y \pi}{L} \right) \right], \quad \kappa S_z = Q_s(y) \]  

(14)

2) The basic steps in numerical experiments

This is the technique invented by F. Bryan (1986). First run the model to a quasi-equilibrium under the Dirichlet conditions. For a model including two hemispheres, the steady state
has two cells, which are symmetric with respect to the equator. Second, diagnose the salt flux \( Q_s(v) \), required for the balance. Third, restart the model from the quasi-equilibrium state and run it under the mixed boundary condition. Sometime in order to find the transition to the other states, an initial perturbation is needed, mostly in form of salinity perturbation at high latitudes. Sometimes, the symmetric state is unstable to very small perturbations, so the solution will drift away without ever adding on any initial perturbation.

(1). Positive salinity anomaly added at high latitude. This leads to a slow evolution toward the one cell solution (pole-pole). The basic mechanism of this instability is the advective feedback discussed by Walin (1985).

(2). Negative salinity anomaly added at high latitude. This leads to a catastrophic change in the meridional circulation. Because water at high latitude becomes too fresh to sink, so the deepwater formation at one polar basin, where the salinity perturbations are added, is suddenly interrupted. As a result, the two-cell circulation pattern suddenly switches to the one-cell circulation. This mechanism is called convective feedback, which has a much short time scale.

(3). The halocline catastrophe in a single-hemisphere model

Marotzke (1989) discussed a model with single hemisphere. After switching into the mixed boundary condition and applying negative salinity perturbation at high latitudes, he found the deepwater formation was suddenly interrupted, and the polar halocline catastrophe set in. The model shifted to a mode that is saline-controlled. In fact, deep water in formed at equator and sink. This mode of circulation is quite slow compared with the thermal mode with sinking at high latitude.

The most interesting aspect of this mode is its instability under the high latitude convective overturning. After many thousands of years, deep water in the polar ocean becomes warm and salty due to slow mixing process. The cold and fresh water lie on top of the warm and salty water at high latitude is potentially very unstable. At certain stage, small perturbations can intricate catastrophic overturning, during which the deep water come to the surface where it looses its heat constant quickly, but retains its salinity. As the result, the water column becomes gravitationally unstable and strong overturning set in. This is most energetic phase of the circulation, which is called $flushing$. At the end of flushing, the polar halocline gradually set in, and the whole cycle repeat.

Note that, the precondition for flushing requires warm and salty deep water in the polar ocean, which is definitely non-existing for the present-day oceans. Whether such phenomenon happened in the past remain to be examined very carefully.

B. The Zonally-averaged model by Wright and Stoker (1991)

As a major improvement over the model by Marotzke et al (1988), Wright and Stoker introduced a parameterization of the zonal pressure gradient term. In Cartesian coordinates, their argument is the following: Assuming semi-geostrophy

\[
f v = \frac{p_e - p_w}{\rho_0 D} \quad \text{(15a)}
\]

\[
f u = \frac{1}{\rho_0} p_y - rv \quad \text{(15b)}
\]

Eliminating \( \vec{v} \), one obtains

\[
\frac{p_e - p_w}{D} = \frac{1}{r} \left( 1 - \frac{u}{u_G} \right) \frac{\partial p}{\partial y} = -\varepsilon_i \frac{\partial p}{\partial y} \quad \text{(16)}
\]

where
\[-u_o = -\frac{1}{\rho_0 f} \frac{\partial \bar{p}}{\partial y}\]  

is the zonally-averaged geostrophic component of the total zonal velocity. Substituting (16) back to (15a), we obtain

\[-v = \frac{\varepsilon_1}{f \rho_0} \frac{\partial \bar{p}}{\partial y}\]  

(18)

Notice that this relation is similar to the assumptions used in the box model that the meridional velocity is proportional to the meridional pressure gradient, but it is different from the parameterization used by Marotzke et al. (1988). The coefficient \(\varepsilon_1\) is diagnosed by averaged data collected from three-dimensional OGCM, it varies from 0.1 to 0.3, depending on the exact model formulation. With this parameterization, their model can provide much more accurate solutions that look almost like the solutions obtained from three-dimensional model by zonal average. Their model has been used to study the global oceans circulation and simulating the Younger Dryas event, with very encouraging results.

**C. Limitation of box models**

1. Most box models have only very few boxes, so they cannot really simulate the details of the general circulation. Although we can construct box models with many boxes, nobody wants to do this because it becomes much easy to construct model based on partial differential equations and finite difference.

2. Numerical diffusion due to upwind scheme.

3. Excluding the Coriolis force. In fact, a 3x3 barotropic box model based on the C-grid can include the Coriolis force and be used to demonstrate the \(\beta\) -effect, i.e., there will be the western intensification in the model. Basically, the two rows to the east play the role of ocean interior, and the western row plays the role of western boundary layer. However, the model is so highly truncated, parameters used in the models are far away from the realistic region.

**D. The limitation of the 2-D models**

1. Difficult to include wind stress and the associated horizontal advection.

2. The velocity is only diagnostic.

3. This is zonally averaged model, so anything associated with the horizontal gyre is not explicitly simulated, thus, the model cannot be used for simulating the decadal/inter-decadal variability of the thermohaline circulation that is essentially a three-dimensional phenomenon.

5. Thermohaline variability

**A. A heat-salt oscillator (Welander, 1982)**

The model consists of a well-mixed layer, forced from above by a heat flux \(k_T(T_a - T)\) and salt flux \(\kappa_s(S_a - S)\), Fig. 14. The turbulent flux between the mixed layer and the water below, with the transition flux, \(\kappa = \kappa(\rho)\), depending on the density difference, Fig. 15. Thus, the basic equations are

\[\dot{T} = k_T(T_a - T) - \kappa(\rho)T\]  

(19)

\[\dot{S} = \kappa_s(S_a - S) - \kappa(\rho)S\]  

(20)

\[\rho = -\alpha T + \gamma S\]  

(21)
where we set $T_0 = S_0 = \rho_0 = 0$. The essential assumption is that $\kappa(\rho)$ is a positive function, monotonically increasing with $\rho$.

The steady state of the system is

$$\bar{T} = \frac{\kappa_T T_A}{\kappa_T + \kappa(\rho)}, \quad \bar{S} = \frac{\kappa_S S_A}{\kappa_S + \kappa(\rho)} \quad (23)$$

$$\bar{\rho} = -a\bar{T} + \gamma \bar{S} = -\frac{a\kappa_T T_A}{\kappa_T + \kappa(\rho)} + \frac{\gamma \kappa_S S_A}{\kappa_S + \kappa(\rho)} \quad (24)$$

A simple case of oscillator is obtained by assuming

$$\kappa = \kappa_0 \text{ for } \rho \leq -\varepsilon$$

$$\kappa = \kappa_1, \text{ for } \rho > -\varepsilon \quad (25)$$

where $\kappa_0$ is small, $\kappa_1$ is large. There are two "attractors" in the phase plane. However, the system can never reach these attractors. As soon as the system across the diagonal in the phase plane, the system is attracted by the attractor on the other side of the diagonal, Fig. 16; thus, the system can never reach a steady state, instead there is a limit cycle, Fig. 17.

Fig. 16. The basic model. The values $T_A$ and $S_A$ are the effective temperature and salinity external forcing; $T_0$ and $S_0$ are the temperature and salinity of the deep reservoir; and $\kappa_T$, $\kappa_S$ and $\kappa$ are the coefficients in assumed Newtonian-type transfer laws. (Welander, 1982)
Fig. 17. Examples of possible forms for $F(\rho)$ in a case where $\kappa(\rho)$ increases monotonically. In the case (a) and (b) there exists one steady state, in the case (c), three steady states. (Welander, 1982).

Fig. 18. Schematic picture of the flip-flop model in a T-S plane. The points 1 and 2 represent the non-existing steady states which act as "attractors" when the solution point is in the half-plane $\rho \leq -\varepsilon$ and $\rho > -\varepsilon$, respectively. The slop dS/dT has a discontinuity at the lines $\rho = -\varepsilon$. (Welander, 1982)
Welander's model is an idealization of the convective overturning in the more complicated three-dimensional OGCM; thus, convective adjustment in OGCM is an important factor in OGCM. Of course, the primarily reason for any thermohaline variability is the nonlinear feedback between different components of the system. Since the system is so complicated, it will take a long time before we can sort out all different types of feedback mechanisms and understand their role in controlling the climate.

Stommel (1987) independently formulated a thermohaline oscillator. Stommel arranged the two boxes horizontally. He assumed that the exchange rate between the boxes becomes very large when the density between them is very small, which presumably mimics isopycnal mixing. His solution poses a limit cycle rather similar to the Welander's model.

### B. Decadal timescale variability

Decadal variability of the thermohaline circulation is now a hot topic because its close relation with the climate variability on decadal time scale. The loop oscillation has been explored as one of the possible mechanisms. The circulation's sensitivity to the virtual salt flux pattern has been examined by Weaver et al. (1991).
Fig. 20. Decadal variability in a single-hemisphere ocean model forced by virtual salt flux. Left: a) Restoring boundary condition on salinity used in the three experiments (WSBC, CBC1, and CBC2); b) Density, $\sigma$, as defined by the restoring temperature and salinity; c) Zonally averaged freshwater flux in (m/yr) as diagnosed from the steady states of the WSBC, CBC1 and CBC2 experiments under restoring boundary conditions. Right: a) Basin-averaged net surface heat flux ($W/m^2$) for the WSBC integration under mixed boundary conditions; b) Power spectral density (using a 256-point fast Fourier transform) of the surface heat flux shown in (a) for the first 8,214 years of integration of the WSBC experiment. The frequency (x-axis) corresponds to the number of cycles over the 8,214 years of integration, Weaver et al. (1991).

C. Diffusive timescale variability -Flushing

Marotzke (1989) found the flushing phenomenon. A model is run to quasi-equilibrium under the relaxation condition for both salinity and temperature with no wind stress. After diagnosing the virtual salt flux, the model is continued to run under the mixed boundary conditions. By adding a small (1 psu) fresh perturbation to the high latitude, the polar halocline catastrophe set in. The system slowly evolved into a state with very weak equatorial downwelling (it is verse of the thermal mode). Within several thousand years the horizontal diffusive process gives rise to warm and salty abyssal water in the whole basin. At high latitude cold and fresh water overlay the warm and salty water, with a relatively weak stratification.
This structure is potentially unstable. Imagine a small water parcel move in the vertical
direction, bringing the warm and salty deep water to the surface. Near the surface water parcel
looses its heat quickly due to the strong thermal relaxation; however, there is no such strong
relaxation for the salinity.

As a result of this cooling, this water parcel becomes heavier than the environment and
sinks to the bottom. The potential energy released from this process will feed energy to the
convective overturning. Consequently, the system will undergo a very violent phase, called flush,
till the potentially storage energy is all consumed. The system is in a quasi-thermal mode during the
flushing period, but it returns to the quasi-haline mode after the flushing. This process can be
described quantitatively by introducing the Diabatically Potentially Available Energy Index
(Huang, 1994).

Fig. 21. A schematic diagram illustrating the meaning of diabatic available potential energy index
(DAPEI) released during the flipping-cooling-overturning process. For convenience, assume
$\alpha = \beta = 1$ and $T$, $S$ and energy are in nondimensional units, Huang (1994).

\[ DAPEI = E_3 - E_4 - (E_2 - E_1) \]
Fig. 22. Time evolution of the total kinetic energy (TKE), diabatically available potential energy index (DAPEI), available potential energy (APE), and buoyancy frequency over several periods of flushing, Huang (1994).

Fig. 23. Time evolution of thermohaline circulation in a single-hemisphere model ocean forced by wind stress, thermal relaxation, and freshwater flux, Huang (1994).

An interesting application of this idea is the deep ocean anoxia in Late Permian (Zhang et al., 2001). As the models, including a three-box model and an oceanic general circulation model,
indicated, the oceanic circulation at that time may be switched between a long-time haline mode and a short-burst of thermal mode, with a period of approximately 3330 years.

Figure 4. Possible thermal mode Late Permian ocean circulation. (a) Horizontal stream function (in Sv) of the Late Permian "thermal mode", revealing a superposition of the wind-driven circulation (Sverdrupian gyres) and the thermohaline circulation looping through the Tethys Sea. (b) Overturning stream function (in Sv) of the Late Permian thermal mode. (c) Sea surface temperature (SST, °C). (d) Meridional ocean heat transport (PW).
Fig. 24. Possible thermal mode and haline mode simulated from a numerical mode. Note that thermal mode has a much faster circulation and large poleward heat flux than that of haline mode (Zhang et al., 2001).

Reference: