7.3 SELF-SUSTAINED THERMOHALINE OSCILLATIONS IN PALEO OCEANS

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1. INTRODUCTION

In past climates the thermohaline circulation (THC) of the ocean may have been quite different from that of today. Paleoclimatic records (e.g. Railsback et al., 1990) suggest that in warm periods of Earth history, warm salty waters may have ventilated the ocean in low latitudes (haline mode circulation), rather than cold water from the pole as in today's climate (thermal mode circulation). Such different ocean circulations would have very different implications for climates and biogeochemical cycles. Self sustained thermohaline oscillations have been found in ocean general circulation model (OGCM) (Marotzke, 1989; Wright and Stocker, 1991; Weaver et al., 1993; Huang 1994). In our study of possible modes of the late Permian ocean circulation (Zhang et al., 2001), we found that the haline mode (HM) was inherently unstable for fixed external forcing, and periodically switched to a transient thermal mode (TM) in which deep water formed near the pole (the so-called flushing event (Marotzke, 1989)). Such internal thermohaline oscillations may have significant implications for understanding the paleo record.

In this study, we construct a 3-box model incorporating a novel representation of convection. It combines the Stommel-type box model (Stommel, 1961) with the Welander-type convection model (Welander, 1982). The model captures the main character and essential physics of the thermohaline oscillation and the instability exhibited in many OGCMs. It illustrates why the steady HM becomes unstable in a certain range of freshwater forcing and vertical diffusivity amplitudes.

2. SELF-SUSTAINED THERMOHALINE OSCILLATION IN OGCM OF LATE PERMIAN OCEAN

We can model thermohaline oscillations with a global high resolution OGCM (MITgcm, Marshall et al. 1997 a,b) configured with late Permian bathymetry under the mixed boundary condition. Details are published in Zhang et al. (2001). The late Permian (250 Myrs ago) had a warm equable climate (Taylor et al., 1992), in which a HM is more likely to occur. We obtained a HM with a freshwater flux maximum ~1.3 m/yr, (somewhat, but not significantly, higher than in today's climate) and a vertical diffusivity 3×10^{-5}m^2/s.

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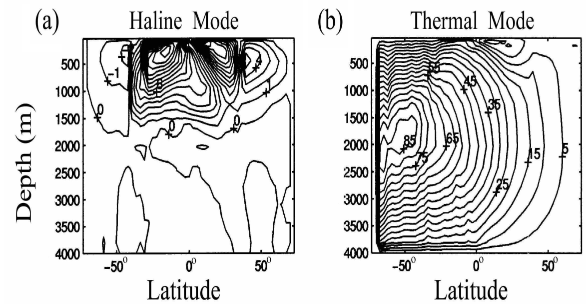

Figure 1: Overturning stream function. (a) HM, (b) TM.

Fig. 1(a) shows the typical overturning stream function of the quasi-steady weak and shallow HM: warm and salty intermediate water is formed in the subtropics, and upwells in polar region and tropics. Fig. 1(b) shows the typical overturning stream function of the transient strong and deep TM: deep water is formed in southern polar regions, upwells in the tropics and the northern hemisphere.

To diagnose the properties of the oscillation in the OGCM, we divided the southern ocean into three boxes: low latitude surface (14°S – 36.6°S, 0 – 50m); high latitude surface (36.6°S – 70.3°S, 0 – 50m) and deep ocean (14°S to 70.3°S, 50 – 4000m). Fig. 2 shows the time series of mean temperature $\widehat{T}_i$ and salinity $\widehat{S}_i$ (i=l,h,d) in each region of the OGCM for about 8000 years. We find an oscillation with a with period of about 3300 years. When $\widehat{T}_d$ is increasing, it corresponds to the long term quasi-steady HM; when $\widehat{T}_d$ begins to drop, we see the onset of convection and a strong, transient TM.

The switch from the quasi-steady HM to the transient TM is triggered by intense local convection in polar regions, induced by meridional transports of warm salty surface water. Fig. 3 shows the time series of the non-dimensional difference between high latitude surface density $\widehat{\rho}_h$ and deep ocean density $\widehat{\rho}_d$: $\Delta\rho_{hd}$. During the unsteady HM, $\Delta\rho_{hd}$ gradually increases, until it reaches a certain threshold, i.e. $\Delta\rho_{hd} = \varepsilon \approx -1.1$. Suddenly strong polar convection starts, $\Delta\rho_{hd}$ jumps to a very high value and the circulation switches to the transient TM. Polar convection can exist even though

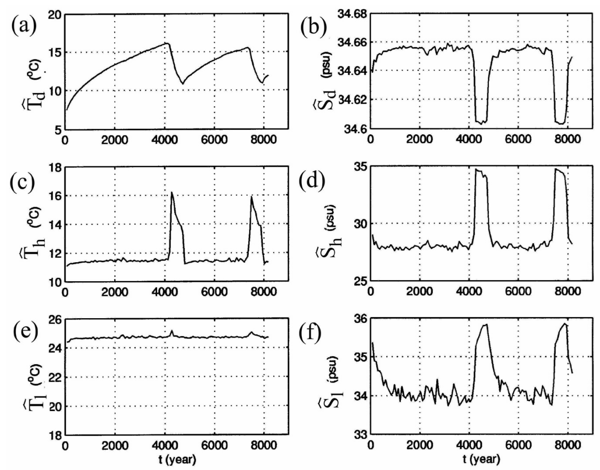

Figure 2: Time series in each diagnosed box of the OGCM. (a) $\widehat{T}_d$,(b) $\widehat{S}_d$, (c) $\widehat{T}_h$,(d) $\widehat{S}_h$, (e) $\widehat{T}_l$, (f) $\widehat{S}_l$.

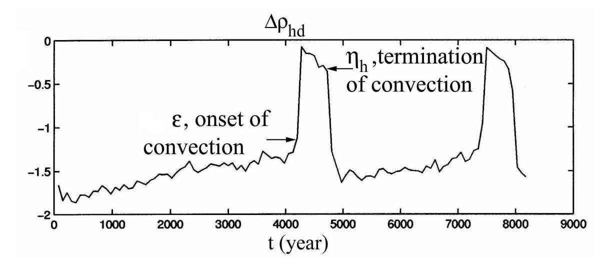

Figure 3: Time series of $\Delta\rho_{hd}$ from the OGCM.

the mean density structure is stable, i.e. $\Delta\rho_{hd} = \varepsilon < 0$.

During the transient TM, the mean trend of $\Delta\rho_{hd}$ (Fig. 3) decreases, since the deep ocean density keeps increasing due to polar convection. Finally when the freshwater forcing prevails, the surface density of the convective region becomes less than the deep ocean density, i.e. $\Delta\rho_{hd}$ reaches the threshold for the termination of polar convection: $\Delta\rho_{hd} = \eta_h \approx -0.3$ (Fig. 3), the convection stops and the circulation returns to the HM.

3. SELF-SUSTAINED THERMOHALINE OSCILLATION IN A SIMPLE BOX MODEL

To better understand the mechanism of the thermohaline oscillation, we developed a 3-box model inspired by the above OGCM results. It includes a low latitude surface box, a high latitude surface box, and a deep ocean box, below the surface (Fig. 4). $\widehat{K}$ is the horizontal diffusivity between surface boxes. $\widehat{M}_l$, $\widehat{M}_h$

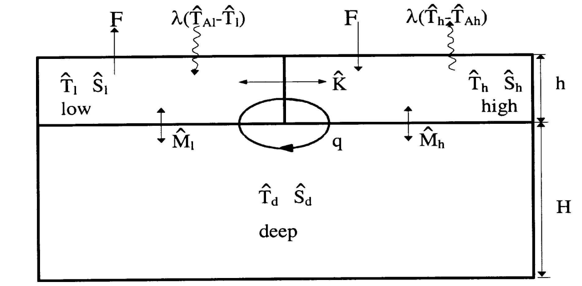

Figure 4: Schematic diagram of the 3-box model.

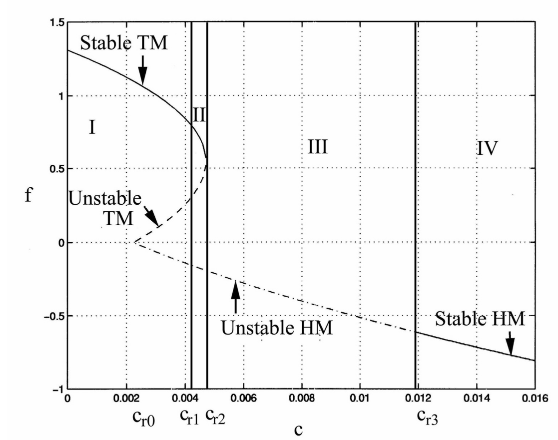

Figure 5: Bifurcation diagram on the f-c plane.

are the vertical diffusivities at low and high latitudes respectively. They depend on the vertical density difference. q is the overtuning circulation strength. The dynamic equations for $\widehat{T}$, $\widehat{S}$ in each box are based on an upstream differencing scheme. We can solve for the steady state and time-dependent solutions of the system.

Fig. 5 shows the steady TM and HM non-dimensional overturning stream function f as a function of the non-dimensional freshwater forcing c when all other parameters are fixed ($f = 1$ corresponds to $q = 21Sv$; $c = 0.005$ corresponds to a mean freshwater flux of $0.5\ m/yr$). Similar to Stommel's box model (1961), there are three branches of steady solutions on the $f - c$ plane (Fig. 5): The strong stable steady TM ($f > 0$, thin solid line), the weak unstable steady TM ($f > 0$, dashed line), the steady HM ($f < 0$). With convective adjustment and localized polar convection, the steady HM could be unstable and four different regions appear on the bifurcation diagram (Fig. 5).

In region III ($c_{r3} \geq c \geq c_{r2}$), no stable steady so-

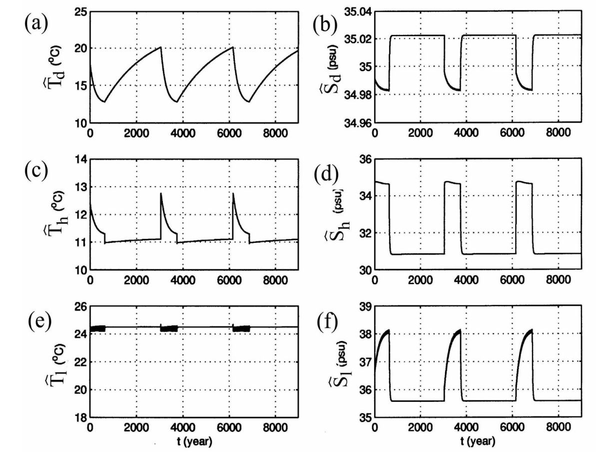

Figure 6: Time series in each box of the 3-box model. (a) $\widehat{T}_d$,(b) $\widehat{S}_d$, (c) $\widehat{T}_h$,(d) $\widehat{S}_h$, (e) $\widehat{T}_l$, (f) $\widehat{S}_l$.

lution exists, the system has to oscillate: the steady HM is the only steady solution, the system evolves towards it until $\Delta\rho_{hd} = \varepsilon$ is satisfied; then suddenly polar convection starts, the system jumps away to the transient TM. Since no steady TM exists in region III, the freshwater gradually becomes dominant until polar convection terminates, the system switches back to the quasi-steady HM and once again evolves to the steady HM, completing the limit cycle. In region IV ($c > c_{r3}$), the steady HM always satisfies $\Delta\rho_{hd} < \varepsilon$, and so no limit cycle exists. In region I ($c_{r1} \geq c$), when $c_{r0} > c$, only one globally stable steady TM exists; when $c_{r1} \geq c \geq c_{r0}$, the basin of attraction of the stable steady TM is big enough that when the unstable HM switches to the transient TM, it will be attracted by the stable steady TM and stay there forever. In region II ($c_{r2} > c > c_{r1}$), the basin of attraction of the stable steady TM is small enough that if the initial state is close to the stable steady TM, the system will evolve to it; if the initial state is close to the steady HM, then the system will evolve to the limit cycle and keep oscillating.

In today's climate, $c < c_{r2}$, and we observe a stable steady TM circulation; and it is perhaps difficult to reach region III. But we find that c_{r2} decreases with the polar-equator air temperature difference. For warm equable climates, it maybe easier to reach the oscillatory solutions in region III.

The time dependent solutions can be solved by numerical integration, and convective adjustment employed at each time step. By choosing $c = 0.0065$ (mean freshwater flux $\sim 0.65m/yr$), background vertical mixing $\widehat{M} = 3.3 \times 10^{-5} m^2/s$, similar to that used in the OGCM and other fixed parameters as in Fig. 5,

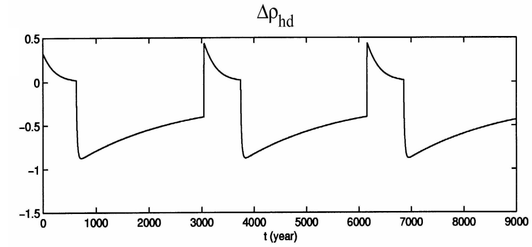

Figure 7: Time series of $\Delta\rho_{hd}$ from the 3-box model.

the system is within the window on the f-c plane where oscillations are possible. We obtain an oscillatory solution with a period of about 3000 years (Fig. 6,7), similar to the OGCM results (Fig. 2, 3).

5. SUMMARY AND DISCUSSION

Using an idealized box model, we have elucidated the underlying mechanisms of haline - thermal mode switching which is a property of OGCMs in certain parameter regimes and perhaps of the paleo oceans. Using steady solutions, and time-dependent integrations of the box model, we are able to fully explore the circulation and stability properties of the system over a wide range of parameter space. The simple model illustrates the inherent instability of the haline mode and the importance of polar thermal convection in the flip mechanism. Motivated by our OGCM results, the localized nature of convection is parameterized in the box model through a threshold criterion, ε, so that polar convection can occur even though the large-scale, mean density structure is statically stable. Oscillatory solutions are found even with a linear equation of state.

On the bifurcation diagram of overturing strength and freshwater forcing, there exists a window in which the steady HM is unstable, and thermohaline oscillations are possible. When the freshwater forcing exceeds an upper limit, only stable HMs exist. For freshwater forcing below a lower limit, only stable TMs exist. Within a broad window of freshwater forcing the HM oscillates. It is notable that it is much easier to switch into the oscillatory haline mode during warm equable climates.

Unfortunately the resolution of most geological records for such warm paleoclimates cannot resolve centuries-millennia timescales, and the long periods of HM circulation might dominate the oscillation. It would be very hard to observe such short periods in those records even if they occurred. During glacial periods, the meridional temperature gradient is stronger and the amplitude of freshwater forcing beyond which no steady TM exists would be larger. On the other hand, the freshwater forcing can be increased by ice melting in polar region and a similar oscillation might be a possibility.

Both the OGCM and box model studies here are forced with prescribed distributions of surface atmospheric temperature and freshwater fluxes. Since the atmosphere provides negative feedback to the air-sea heatflux, we speculate that such oscillating solutions will exist in a coupled atmosphere-ocean system and this is the focus of ongoing study.

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