

Equilibrium thermal response timescale of global oceans

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[1] The equilibrium response timescale of global oceans is estimated in a fully coupled climate model. In general, the equilibrium timescale increases with depth, except in the polar region. The timescale is approximately 200 years for the ocean for depths above 1 km, and it increases to 1500 years at a depth of 3 km. A layer with a rapid timescale change, referred to as a *temporacline*, is located at a depth of 1.5–2 km, which is analogous to the permanent thermocline in the ocean. The equilibrium timescale varies with the sign of the change in radiative forcing. The ocean response to surface cooling could be twice as fast as the surface warming because of enhanced vertical mixing, convection and overturning circulation. However, this phenomenon only occurs below the Atlantic *temporacline*. For the Atlantic upper ocean, the timescale is longer in the cooling case because of the readjustment of the upper ocean to the enhanced Atlantic overturning circulation. In the Pacific, the timescale change in the warming and cooling cases is not as significant as in the Atlantic because of the lack of deep convection. **Citation:** Yang, H., and J. Zhu (2011), Equilibrium thermal response timescale of global oceans, *Geophys. Res. Lett.*, 38, L14711, doi:10.1029/2011GL048076.

1. Introduction

[2] Because of their enormous inertia, the equilibrium response time for oceans can be much longer than 1000 years. The climate response timescale has therefore been less studied in the past several decades because powerful computers and long integrations of complex climate models are needed. Climate researchers are interested in two questions: how long the climate response timescales are and whether the response of timescales depends on the sign of the change in radiative forcing. *Manabe et al.* [1991] and *Stouffer* [2004, hereinafter S04] investigated these questions using coupled models. *Manabe et al.* [1991] focused on the transient response of the model to heating and cooling perturbations, while S04 focused more on the equilibrium response. They all found that the heat anomalies penetrate to a much deeper depth in response to surface cooling than in response to surface warming. In a warming world, the heating from the ocean surface makes the ocean more stable, isolating the deeper waters from the surface. In a cooling world, the ocean tends to be unstable, which enhances mixing between the surface and deeper waters. S04 provided detailed structures of the global ocean response timescale and demonstrated the significant interhemispheric asymmetry and the asymmetry of

the response timescale to the forcing signs. The climate response timescale is usually studied using numerical models, and therefore, its robustness remains to be clarified. This problem motivated us to revisit these questions using a different coupled model. By comparison with previous works, we can quantify to what extent the response timescale is model dependent.

[3] The Fast Ocean-Atmosphere Model (FOAM) [*Jacob*, 1997] is used in this work. The FOAM is computationally efficient and has been used to simulate the global climate in the past, present and future [*Yang and Liu*, 2005; *Wu et al.*, 2003]. To estimate the climate response timescale, we performed two long integrations, warming and cooling experiments, in which the model is forced by increased and decreased CO₂ for 2000 years. Starting from a present-day control run with the 1990 CO₂ level of 355 ppmv, the warming (cooling) experiment is forced with a 1% yr⁻¹ increase (decrease) in CO₂ for 70 years and then continues the integration for 1930 additional years forced by the doubled (halved) CO₂. For comparison, a parallel control run is also integrated for the same length of time. By the end of the integration, the climate is nearly in quasi-equilibrium. In this work, the responses averaged over the last 100 years of the integration are treated as the equilibrium climate sensitivity (ECS). The time required to reach 70% of the ECS is defined as the equilibrium response time (ERT) (S04). The ERT in our model is qualitatively consistent with that given by S04. Our results show that the ocean has a region with an abrupt increase in response timescale, which is referred to as a “temporal-cline” (hereafter, *temporacline*), mimicking the definition of the main thermocline. The ERT above and below the *temporacline* are approximately 200 and 2000 years, respectively. The effect of the forcing sign on the ERT is significant, which is illustrated clearly by the vertical displacement of the *temporacline*. The mechanisms of the climate response are also discussed.

2. Timescale of Oceans

2.1. Global Timescale

[4] The temporal evolution of the global temperature shows that a quasi-equilibrium can be roughly attained in approximately 2000 years for the upper 3 km of the ocean (Figure 1a). Under the doubled CO₂ forcing, the equilibrium surface air temperature (SAT) and sea surface temperature (SST) sensitivities are approximately 2°C and 1.5°C, respectively, values that are in line with the IPCC-AR4 assessment [*Intergovernmental Panel on Climate Change*, 2007]. The equilibrium sensitivities are approximately 1°C for the 40–400 m ocean depths and 0.8°C for the 1–3 km ocean depths, respectively. For the deep and bottom ocean, the response is still weak, even after the 2000-year integration. It is clear that the temperature changes for the doubled CO₂ and halved CO₂ scenarios are nearly symmetric because the changes

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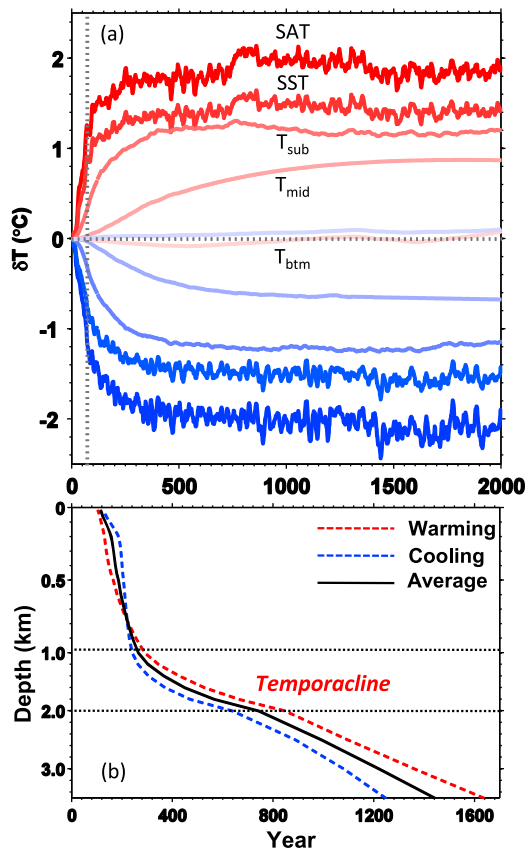


Figure 1. (a) Evolution of the anomalous global annual mean surface air temperature (SAT), sea surface temperature (SST), subsurface temperature T_{sub} (averaged between 40–1000 m), intermediate water temperature T_{mid} (averaged between 1 km–3 km) and bottom water temperature T_{btm} (averaged between 3 km–5 km) in warming (red) and cooling (blue) experiments (unit: °C). The anomalous temperature is obtained by removing the corresponding value from control run. (b) Vertical profile of equilibrium response time (ERT) of global mean ocean temperature. The red dashed, blue dashed and black solid lines are for warming, cooling experiments and their average, respectively.

in the radiation forcing caused by the CO_2 change are symmetric.

[5] The ERT is approximately 200 years for the upper 1 km of ocean, and it increases almost linearly from 400 years at a depth of 1.5 km to 1500 years at a depth of 3.5 km (Figure 1b). It is interesting to see a triplex structure in the vertical profile of the timescale. The ERT is more or less the same in the upper 1 km. The ERT increases rapidly from 200 to 800 years at 1–2 km and then increases modestly from 800 to 1500 years below this depth. This structure is understandable when considering the global ocean temperature profile. The rapid time change zone at 1–2 km is analogous to the main thermocline in the ocean, although the depth of the latter is generally shallower. Here, the rapid timescale change layer is the *temporacline*. The 200 years in the upper ocean corresponds to the wind-driven layer, while the longer timescale in the deep ocean corresponds to the layer driven by the thermohaline processes at high latitudes. The *temporacline* defined in this

work is a useful and convenient concept to describe the ocean response timescale. The ERT discussed here is obtained by averaging the corresponding timescale in the warming and cooling experiments. Because the changes in radiation forcing can change the ocean circulation or dissipation processes and, thus, the oceanic ERT, the average between them is more appropriate for representing the internal ocean timescale.

[6] More details on the global ocean ERT are shown in Figure 2. First, the structure of the ERT shows flat contours, except in the polar latitudes (Figures 2a and 2e). In most regions, the ERT increases monotonically with depth because the ocean stratification is generally stable, and the surface signal affects the lower ocean gradually through quasi-horizontal subduction and vertical mixing. In polar regions, the surface thermohaline processes in the north and the strong Ekman pumping in the south tend to destabilize the upper ocean and retard the surface ocean from reaching equilibrium. The ERT is far beyond 100 years, even for the surface ocean, and it exceeds 400 years for the subsurface ocean above 1 km (Figures 2a and 2e), which is much longer than the 200-year timescale in the low-mid latitudes. Second, the *temporacline* is clear in all oceans, and its location is consistent with that of the permanent thermocline. It can be seen that the ERT above the *temporacline* is primarily determined by the Pacific (Figures 2a and 2i), while that below the *temporacline* is mainly determined by the Atlantic (Figures 2a and 2e). The *temporacline* structure itself is determined by the Atlantic.

2.2. Different Timescales in Warming and Cooling Experiments

[7] Significant differences in the ocean ERT are found between the global warming and cooling scenarios (Figure 2). The lower (upper) ocean reaches equilibrium faster (slower) in the cooling case than in the warming case. Although the climate sensitivity (e.g., the global mean temperature change) is nearly symmetric in the warming and cooling scenarios (Figure 1a), the speeds of ocean responses are remarkably different. The *temporacline* is shallower and more intense in the warming experiment (Figures 2b and 2f) than in the cooling experiment (Figures 2c and 2g). For the ocean below 1.5 km, the ERT in the cooling experiment is approximately 500–800 years shorter than in the warming experiment (Figures 2d and 2h). The *temporacline* also shifted downward because of the enhanced vertical mixing and convection. This effect occurs primarily in the Atlantic. In the Pacific, the ERT does not change much, and the *temporacline* shifts downward slightly because of the weak thermohaline circulation and ventilation (Figures 2i–2l). It is worth noting that for the upper ocean above the *temporacline*, the ERT could be 200–400 years longer in the cooling case (Figures 2d and 2h), which is opposite that of the lower ocean. This effect is particularly clear in the northern high latitudes. As a consequence of the destabilization effect of surface cooling, the initially rapidly adjusting upper ocean has to readjust to the changed thermohaline circulation, which prolongs the ERT of the upper ocean in the North Atlantic.

2.3. Different Timescales in the Pacific and Atlantic

[8] There are also significant differences in the ERT in different basins. Generally, the upper layer in the high

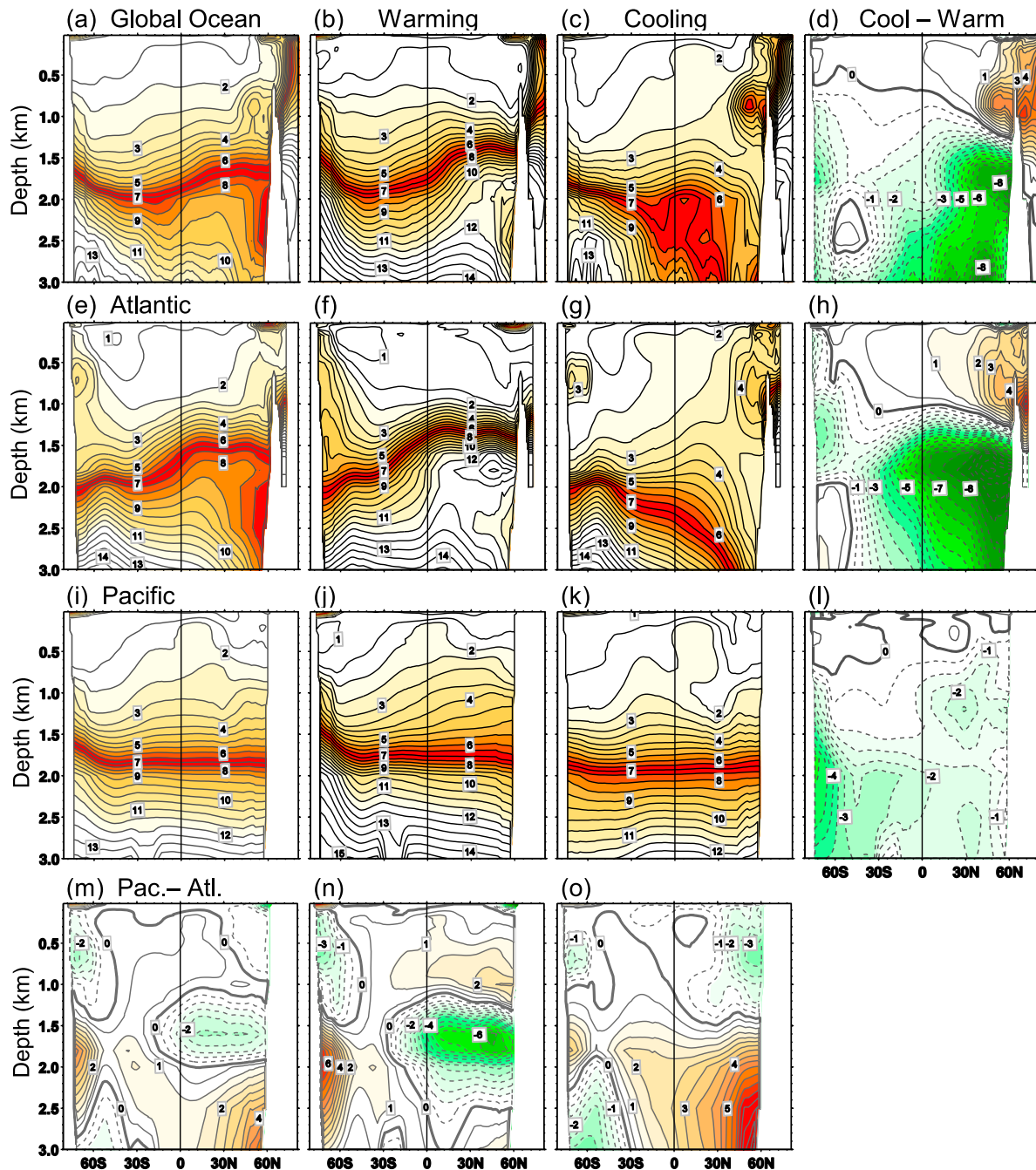


Figure 2. (a–d) ERT of zonal mean ocean temperature for global ocean, (e–h) Atlantic and (i–l) Pacific (unit: 100-year, contour interval: 0.5). (m–o) The ERT difference between Pacific and Atlantic. Figures 2a, 2e, 2i, and 2m (Figures 2d, 2h, and 2l) show the average (difference) of the ERT in cooling (Figures 2c, 2g, 2k, and 2o) and warming (Figures 2b, 2f, 2j, and 2n) experiments.

latitudes in the Pacific reaches equilibrium faster than the upper layer in the Atlantic (Figure 2m) because the Pacific is dominated by wind-driven circulation. It is noted that the ERT changes in the Pacific and Atlantic are out of phase in the warming and cooling scenarios. For the ocean above the *temporacline*, the Atlantic ERT is shorter (longer) than the Pacific ERT in the warming (cooling) experiment. This situation is reversed for the ocean below the *temporacline* (Figures 2n and 2o). The greatest difference in the northern high latitudes: in a warming (cooling) climate,

the lower Atlantic ocean would take a much longer (shorter) time than the Pacific to reach equilibrium, even 600 years longer (shorter), because it is the most affected by the slower (faster) meridional overturning circulation.

[9] Finally, the interhemispheric asymmetry in the ERT in the Atlantic should be noted. The asymmetry is clear for the ocean at the 1 – 3 km depth (Figure 2e) and is enhanced in both the warming and cooling cases. The ERT in the northern Atlantic changes the most because of the dramatic changes in the northern deep water formation processes

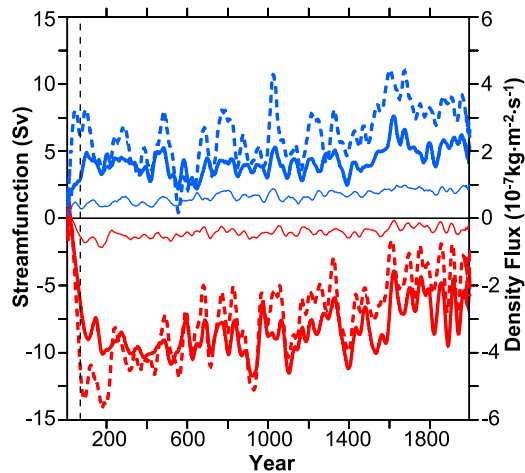


Figure 3. Evolution of anomalous Atlantic Meridional Overturning Circulation (solid, left axis in Sv), thermal density flux (dashed, right axis in $10^{-7} \text{ kg} \cdot \text{m}^{-2} \cdot \text{s}^{-1}$) and saline density flux (thin line, right axis) for warming (red) and cooling (blue) experiments. A 51-year running mean are applied. Total density flux $F_{\rho} = F_{\theta} + F_S$, where $F_{\theta} = -\alpha(Q/c_p)$ and $F_S = \rho(0, \theta)\beta[F_W * S/(1 - S)]$ are thermal and saline density flux, respectively. α , β , ρ , S , θ , c_p , Q and F_W are thermal expansion coefficient (unit: $1/^{\circ}\text{C}$), saline contraction coefficient (unit: $1/\text{psu}$), sea water density, salinity, potential temperature, specific heat at constant pressure, net surface heat flux and fresh water flux, respectively. The streamfunction is obtained by extracting the vertical maximum in region (60°W – 10°E , 50 – 70°N). The density fluxes are the same region averaged.

(Figures 2f and 2g). In contrast, the ERT in the Pacific is nearly interhemispheric symmetric because of the lack of thermohaline circulations (Figures 2i–2k).

2.4. Ocean Circulation Change

[10] The ERT is closely related to ocean mixing, convection and circulation, in which the shallow meridional overturning circulation (or the subtropical cell, STC) in the Pacific and the deep meridional overturning circulation in the Atlantic (AMOC) play important roles [Wunsch and Heimbach, 2008]. Here, we focus on the AMOC. It can be seen that the AMOC changes asymmetrically in the warming and cooling experiments (Figure 3). It is recognized that the AMOC would be weakened in response to increased CO_2 in the short term [Schmittner et al., 2005], but uncertainties exist in its long-term evolution [Manabe and Stouffer, 1994; Rahmstorf and Ganopolski, 1999; Stouffer and Manabe, 2003]. Views of the impact of decreased CO_2 on AMOC diverge greatly [Stouffer and Manabe, 2003; Liu et al., 2005]. In our experiments, the AMOC is weakened (enhanced) in response to increased (decreased) CO_2 , and the magnitude of the AMOC change in the warming experiment is approximately twice that in the cooling experiment (Figure 3), which is dramatically different from the corresponding temperature changes shown in Figure 1a. This result is due to the AMOC in our model being predominantly determined by the local surface density flux in the northern Atlantic high latitudes, while the latter changes asymmetrically in the warming and cooling experiments (Figure 3). The peak correlation between

the AMOC and the local density flux is approximately 0.8, occurring at the latter and leading the former by 2–4 years (figure not shown). The surface density flux consists of the heat flux and the fresh water flux [Schmitt et al., 1989; Shin et al., 2003]. In our experiments, the local surface heat flux change contributes nearly 90% (60%) of the change of the total density flux in the warming (cooling) experiment (Figure 3). The density change induced by the heat flux is the product of the thermal expansion coefficient and the heat flux change. The thermal expansion coefficient, in turn, depends on the temperature, and it increases with an increase in temperature. In the warming and cooling experiments, the heat flux changes are nearly the same. However, because of the higher temperatures in the warming case, the larger thermal expansion effect causes a larger density flux change than in the cooling experiment, which eventually results in the asymmetric changes in the AMOC. This is the mechanism of the large ERT difference for the intermediate and deep oceans between the warming and cooling scenarios.

[11] For the upper ocean, the change in the mixed layer depth (MLD) should be responsible for the ERT change. The surface warming (cooling) makes the upper ocean more stable (unstable), consequently the MLD becomes shallower (deeper), which shortens (lengthens) the upper-ocean ERT. In the Pacific, the overall mean MLD is approximately 150 m, and it decreases (increases) by 15% (20%) in the warming (cooling) experiment (figure not shown). The MLD is shallow, and the effect on the ERT is insignificant (Figure 2l). In the Atlantic, the effect of the MLD can reach a depth of 1 km, which is particularly obvious in the northern high latitudes. Because of the enhanced vertical convection there, the upper ocean takes more than 200 years longer to reach equilibrium in the cooling experiment than in the warming experiment (Figure 2h).

3. Summaries and Discussions

[12] This work quantified the ocean ERT in a fully coupled climate model. The ERT varies widely with latitude, basin, and sign of change in radiative forcing. The ocean response to surface cooling can be twice as fast as the surface warming because of enhanced vertical mixing, convection and overturning circulation, although this only happens in the lower Atlantic ocean. For the upper Atlantic ocean, the ERT is longer in the cooling case because of the readjustment of the upper ocean to the changed AMOC. In the Pacific, the ERT change in the warming and cooling cases is insignificant because of the lack of deep convection.

[13] There are some minor differences between our results and those of S04. The ERT in our model is slightly (approximately 15%) shorter than that in S04 in both the warming and cooling experiments. S04 gives a local maximum timescale of approximately 2 km in the main part of the Pacific and Atlantic, while in our model, the ERT increases monotonically with depth. Our coupled model is significantly different from the model S04 used; specifically, the Antarctic bottom water (AABW) in our model is weak, which implies a weak ventilation of the bottom water. The consequences are twofold: first, the bottom water response to the surface forcing will be extremely long, and second, the AMOC can be very strong, which in turn could result in a short ERT in the intermediate and deep waters. The 200-year timescale for the upper ocean given in our model is consistent with

that estimated in simple one-dimensional advection-diffusion models [Hansen et al., 1985; Dickinson and Schaudt, 1998; Gnanadesikan et al., 2007], which suggests that the mixing scheme and magnitude, particularly for the vertical mixing, might be reasonably represented in our coupled model. The accurate length of the response time depends on the details of the mixing. This work provides reasonable guidance on the climate response timescale.

[14] The long ERT for the deep ocean implies that the climate state, particularly the ocean state at present, is a mixture of the changes to external forcings that have occurred over the past few thousand years (S04). The future climate projection can be affected by past external (mainly radiative) forcings through the initial conditions used to make those projections [Weaver et al., 2000]. In the decadal and longer timescale, the feedback to the atmosphere by the previous changes stored in the ocean can gradually become significant [Held et al., 2010], which implies the future climate prediction would be strongly related to the initial conditions causing the ocean changes. A key suggestion of this work is that for better prediction of future climate changes, further investigations should focus on finding the best technique to initialize coupled climate models (S04).

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