The Freshening of Surface Waters in High Latitudes: Effects on the Thermohaline and Wind-driven Circulations

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Abstract.

The impacts of a freshening of the surface waters in high latitudes on the deep, slow, thermohaline circulation have received enormous attention, especially the possibility of a "shut-down" in the meridional overturning that involves sinking of surface waters in the northern Atlantic. A recent study (Fedorov *et al* 2004) has drawn attention to the effects of a freshening on the other main component of the oceanic circulation - the swift, shallow, wind-driven circulation with which is associated the ventilated thermocline and which varies on decadal timescales. That circulation too involves meridional overturning with surface waters sinking in certain subduction zones, but its critical transition affects mainly the tropics. A freshening can deepen the equatorial thermocline to such a degree that temperatures along the equator become as warm in the eastern part of the basin as they are in the west, so that the tropical zonal SST gradient virtually disappears and permanently warm conditions prevail in the tropics. In a model that has both the wind-driven and thermohaline components of the circulation, which factors determine the relative effects of a freshening on the two components and its impact on climate? Studies using an idealized oceanic GCM to address this question find that the vertical diffusivity of a model is one of the critical parameters that changes the relative strength of the two circulation components and hence their response to a freshening. The spatial structure of the surface freshwater forcing and imposed meridional temperature gradients are other important factors.

1. Introduction.

The surface waters in high latitude oceans are likely to freshen over the next few decades should polar ice caps melt, or should precipitation increase in those regions. Recent data indicate that this has already happened over the last several decades in the Atlantic Ocean (Dickson *et al* 2002, Curry *et al* 2003). What are the possible climate impacts? A vast number of studies have concentrated on the changes in the thermohaline circulation (THC) – a component of the global meridional overturning circulation associated with the surface temperature and salinity gradients and involving the deep ocean. That a freshening in high latitudes can lead to a shut-down of the thermohaline circulation is the topic of many papers (e.g. Manabe and Stouffer 1995, 2000, Rahmstorf 1995, Stocker and Schmittne 1997, Alley *et al* 2003, Seidov and Haupt 2002, 2003 and many others). The impact of such a "shut-down" on the Earth's surface temperatures is prominent mainly over the northern Atlantic and western Europe.

Recently, Fedorov *et al* 2004 demonstrated, by means of an idealized general circulation model configured for the size of the Pacific basin, that a similar freshening can also affect the shallow, wind-driven circulation of the ventilated thermocline and its heat transport from regions of gain (mainly in the upwelling zones of low latitudes) to regions of loss in higher latitudes. A freshening that decreases the surface density gradient between low and high latitudes reduces this poleward heat transport, thus forcing the ocean to gain less heat in order to maintain a balanced heat budget. This is achieved through a deepening of the equatorial thermocline. (A weakened meridional density gradient implies a deeper thermocline. The deeper the thermocline in equatorial upwelling zones is, the less heat the ocean gains.) For a sufficiently strong freshwater forcing, the poleward heat transport from the equatorial region all but vanishes, and permanently warm conditions with absent zonal SST gradient along the equator ("permanent El Niño" in the case of the Pacific, e.g. Fedorov *et al* 2006) prevail in the tropics.

What is the connection between the wind-driven and thermohaline components of the oceanic circulation? Although some studies of the THC give the impression that the Gulf Stream is exclusively one of its features, this current will be absent should the wind stop blowing (to the extent that its volume transport is proportional to the integrated wind stress curl in the interior of the ocean), even though a THC will still be possible. On the other hand, the theories of the ventilated thermocline (e.g. Luyten *et al* 1983, Huang 1988, Pedlosky 1996) are incomplete and require specification of the thermal structure of the ocean along the eastern boundary of a basin. Alternatively, the oceanic stratification in the absence of winds needs to be specified. Presumably that stratification is determined by the thermohaline circulation which therefore is implicit in the theories for the ventilated thermocline. (The latter theories can be regarded as three-dimensional versions of Stommel's 1948 model of the Gulf Stream.) A number of studies looked at particular features of the oceanic circulation when both buoyancy and wind forcing are present (e.g. Veronis 1976, 1988, Luyten and Stommel 1986, deSzoeke 1995), but the overall picture remained incomplete. Boccaletti *et al* (2004) recently attempted to eliminate the indeterminacy of the ventilated thermocline theories, and integrated them with theories for the thermohaline circulation by invoking the constraint of a balanced heat budget for the ocean.

In Fig. 1 the ocean is seen to gain a very large amount of heat in the equatorial upwelling zone. This would lead to a rapid deepening of the thermocline were it not for the currents that transport the warm water poleward so that the heat is lost in higher latitudes, especially where cold, dry continental air flows over the warm Gulf Stream (and the Kuroshio Current in the Pacific). In a state of equilibrium the gain must equal the loss. Should there be a warming of the atmosphere in mid-latitudes that reduces the oceanic heat loss then warm water accumulates in low latitudes so that the thermocline deepens. The winds then fail to bring cold water to the surface at the equator, the gain of heat is reduced, and a balanced heat budget is restored. It follows that the thermal structure depends critically on the fluxes of heat across the ocean surface. If the loss of heat in high latitudes is large then the ocean poleward transport of heat is large, and the gain of heat in low latitudes must be large. Exactly how the ocean gains heat depends on its vertical diffusivity. If the latter parameter is small, then the thermocline is sharp and has to be shallow for the ocean to gain a large amount of heat. If, on the other hand, the diffusivity is large then the thermocline is diffuse and the ocean gains heat by diffusing it downwards into the deep ocean.

How a freshening of the surface waters can cause a "shut-down" of THC, and change the wind-driven circulation, follows from the linearized expression for the equation of state that describes the dependence of density ρ on temperature T and on salinity S:

$$\rho = \rho_0(1 - \alpha T + \beta S) \tag{1}$$

where ρ_0, α and β are constants (in general, α and β are functions of temperature and salinity). The maintenance of warm conditions in low latitudes, and of cold conditions in polar regions drive a circulation. If density should depend on temperature only, then this circulation disappears when the imposed meridional temperature gradient ΔT vanishes. In the presence of salinity variations, the meridional density gradient, and hence the thermal circulation can disappear even in the presence of a meridional temperature gradient. Equation (1) implies that this happens when

$$\Delta\rho = \rho_0 (\alpha\Delta T - \beta\Delta S) \to 0 \quad \text{or} \quad R \to 1 \quad \text{where} \quad R = \beta\Delta S / \alpha\Delta T. \tag{2}$$

Thus, for the case of the thermohaline circulation the limit R -> 1 is known to be associated with a "shut-down" of that circulation (e.g. Zhang *et al* 1999).

For the wind-driven circulation the same limit (R -> 1) leads to the establishment of warm tropical conditions with no zonal SST gradient (Fedorov *et al* 2004). Such singular behavior occurs because the equator-to-pole density difference $\Delta\rho$ at the surface is also a measure of the vertical density gradient (or the vertical stability of the ocean) in low latitudes. (Note that meridional density gradients that control the wind-driven circulation can be effectively different from those controlling the THC.) When the freshwater forcing exceeds a critical value and R becomes greater than unity, the thermocline deepens, its slope in the equatorial plane collapses, the cold equatorial tongue at the surface vanishes, and permanently warm conditions prevail in the tropics ("permanent El Niño" in the case of the Pacific). The collapse of the thermocline eliminates the confined equatorial region where the ocean gains a large amount of heat (in the tropics it is the depth of the thermocline that controls the heat intake by the ocean). As a result, the constraint of a balanced heat budget is no longer satisfied by transporting heat from regions of gain to regions of loss. Rather, the heat budget tends to be balanced more locally, and the heat transport from the equatorial region is significantly reduced. Apparently such a state of affairs can be induced merely by freshening the surface waters in the extra-tropics.

The calculations on changes in the wind-driven circulation, as described in Fedorov *et al* 2004, were conducted with high resolution models for relatively short periods not exceeding 30-60 years. The time scales of the upper ocean adjustment are much shorter than those for the deep ocean, so that by the end of the integrations the upper ocean reaches a quasi-steady state while the deep ocean still undergoes a gradual adjustment. This approach allows many calculations with relatively modest computer resources but does not provide any information about changes in thermohaline circulation. Nor does it give any indications of how the shallow wind-driven circulation can interact with the deep meridional overturning. (The main application for that study was the Pacific basin where the THC does not play a major role.) Would our arguments on the shutdown of the wind-driven circulation change in the presence of an active thermohaline circulation? What are the factors that determine whether a freshening of the surface waters of the Northern Atlantic results in a colder climate for northwestern Europe (in accord with results from studies of the thermohaline circulation), or a deeper equatorial thermocline (in accord with the results of Fedorov *et al* 2004)?

2. The model.

To address the questions raised above we follow the approach used by Zhang, Schmidt and Huang (1999) but extend their model to include the tropics. Those authors confined their attention to the thermohaline circulation and its response to a freshening of surface waters, and systematically studied the effect of changing the thermal diffusivity and of other relevant parameters in a general circulation model in a basin stretching from 60^oN to 5^oN. A large number of calculations with different strengths of the freshwater forcing were conducted. In each experiment the model was run to reach a steady state, unless none was found. By varying freshwater forcing Zhang *et al* observed a classical hysterisis in which increasing freshwater forcing leads to a gradual reduction and then a sudden collapse in the poleward heat transport (with decreasing forcing the system would recover to the original values of the poleward heat transport but at a much slower pace).

In this study the General Circulation Model MOM4 was first used to reproduce the results of Zhang, *et al* 1999. Then we extended the domain to the latitude band 60^oN to 60^oS. In order to speed up the calculations, we use a version of the model with a relatively coarse resolution (3.75^ox3.75^o) as Zhang *et al* did. The width of the basin matches that of the Atlantic. There are 15 levels in the vertical and the bottom at 4500m depth is flat. In many instances a steady or a quasi-steady state can be reached in the course of 4000 years of integration. The zonal wind stress is specified as in Bryan (1987), see the Appendix, and does not change from one experiment to the next. Free-surface boundary conditions rather than the rigid lid (as in Zhang *et al*) are used, which affects the model solutions only slightly. Given the coarse model resolution and simple geometry, the intent is not to reproduce reality in detail, but rather to study the main tendencies for the problem.

The oceanic processes we explore affect sea surface temperatures in the tropics, and hence can have an impact on the global climate, but those processes are entirely different from the ones that cause El Niño and La Niña. Whereas the latter phenomena are associated with an adiabatic redistribution of warm surface waters in the tropical Pacific, the processes to be explored here involve changes in the oceanic heat budget and hence are diabatic. To change the heat content of the ocean, two mechanisms are available: the one is direct and involves changes in the heat flux across the ocean surface; the other is indirect and involves changes in meridional salinity gradients which can also interfere with the oceanic heat budget (see eqs. (3)).

The standard mixed boundary conditions on temperature and salinity satisfy the above requirements:

$$kT_z = -A(T - T^*) \qquad\qquad kS_z = (E\text{-}P)S_0 \qquad\qquad (3)$$

where A=0.6 m/day is a constant (equivalent to 25 $W/m^2/K$ in terms of the appropriate heat flux), z is the vertical coordinate, (E-P) is evaporation minus precipitation and river runoff, S_0 is the mean salinity in the basin, T is the sea surface temperature, and T* is an imposed temperature, and k is the coefficient of vertical diffusivity at the surface. In order to have a realistic equatorial cold tongue in the eastern equatorial part of the basin, in a relatively coarse model, the factor A^{-1} (proportional to the restoring timescale, see Haney 1971) is chosen to be somewhat smaller than commonly used - otherwise the positive surface heat flux in the tropics would overwhelm equatorial upwelling and the cold tongue in the eastern equatorial part of the basin would disappear.

3. Results: "Freshening" experiments.

Here, we describe in details our numerical experiments with an intermediate (presumably realistic) value of the background diffusion coefficient (0.5×10^{-4} m^2/s), and then outline the main differences with low and high diffusion. The methodology of this study is as follows. Firstly, we produce standard initial conditions, used in subsequent perturbation experiments, by running the model for 4000 years until a steady state is reached. The restoring temperature T* for these experiments is taken from the observations along the 30^{o}W longitude (Levitus and Boyer 1994) but made symmetric with respect to the equator (fig. 2a). A small, linearly varying with latitude E-P flux is added to the forcing (this flux is positive in the Northern and negative in the Southern hemisphere, see equation (A.2) of the Appendix). With such boundary conditions the model produces a realistic structure of the Atlantic meridional overturning circulation with preferential sinking in the Northern Hemisphere and the maximum overturning streamfunction of about 17 Sv (figs. 3a and 4a).

Next, we proceed to the perturbation experiments - from one experiment to the next we keep T* the same but vary the E-P flux and explore changes in the relevant characteristics of the ocean circulation and thermal structure after 100 years of integrations. The shape of the E-P forcing for these perturbation experiments (equation (A.3) of the Appendix) is chosen to induce a freshening of surface waters in high latitudes but to conserve mean salinity.

Fig. 2 shows the forcing used in some of the "freshening" experiments and simulated changes in the zonal SST along the equator for five different strengths of the E-P flux, with the freshwater forcing at 55^{o}N corresponding to 0, 0.5, 0.8, 1.1, and 1.7m/year. The calculated meridional overturning for cases 1 and 2 is shown in fig. 3 and 4. Summary plots which include several additional experiments are presented in fig. 5 that combines changes in the equatorial SST gradient, the depth of the tropical thermocline, and the THC volume transport as a function of the freshwater flux.

As in many previous studies, with increasing freshwater forcing the strength of the overturning decreases and it ultimately "breaks down" or collapses (figs. 3b, 4b and 5). At the same time, with increasing freshwater forcing and in accordance with the arguments of the previous section, the equatorial SST gradient weakens and then collapses, which occurs in this set of experiments almost simultaneously with the THC collapse. As further calculations suggest, however, the simultaneous collapse of the SST gradient and the THC in these particular calculations is a mere coincidence – a slightly different combination of the restoring timescale and/or background diffusion would make possible a collapse of only one of the two.

The reduction and then reversal of the equatorial SST gradient (fig 5a) is associated with the reduction in the meridional density gradient (measured from high latitudes to the equator) and with the deepening of the equatorial thermocline as seen in fig. 5b. (Note that Johnson and Marshall (2002) studied changes in the thermocline depth within a reduced-gravity model when a THC strength is varied and also noticed a deepening of the tropical thermocline associated with heat accumulation in the upper ocean. Their model, however, did not have mechanisms allowing for the equatorial upwelling or wind controls of the equatorial thermocline.)

Fig. 4 provides a detailed view of the overturning circulation for cases 1 and 2 for the upper 500m of the ocean, which emphasizes the role of the shallow wind-driven cell effecting a significant fraction of the heat transport from the tropics. The strength of this component of the overturning circulation changes only slightly, from one experiment to the next; however, its impact on the tropics and its relative contribution to the heat transport varies significantly.

Numerous studies have established that the strength of ocean mixing is one of the key factors that determine the strength of the THC in ocean models (e.g. Park and Bryan 2000). To explore this issue we repeated most of the experiments of this study with large ($5\text{x}10^{-4}$ m^2/s) and small ($0.05\text{x}10^{-4}$ m^2/s) values of the coefficient of vertical diffusion. In agreement with previous studies, our calculations indicate that in a model with large values of diffusivity the thermohaline circulation is dominant. The wind-driven circulation gains in importance as diffusivity decreases, dominating the circulation for low values of the diffusion. The strength of the shallow overturning associated with the wind-driven component of the ocean circulation remains largely unchanged; rather, the strength of the THC varies. In the experiments with low values of diffusivity ($0.05x10^{-4}$ m^2/s) the thermohaline circulation is all but nonexistent, and the poleward heat transport is maintained almost exclusively by the wind-driven circulation.

A series of perturbation "freshening" experiments was conducted for the low and high values of the diffusion coefficient. In the experiments with small diffusion the zonal SST gradient along the equator collapses at a relatively low magnitude of the freshening (0.8m/s). In the experiments with large diffusion, there is a vigorous thermohaline circulation, and a larger freshwater forcing (2.5m/year) is needed to collapse the equatorial SST gradient. Even a stronger forcing would be necessary to collapse the THC. The results of these experiments on the role of ocean diffusion are combined in Table 1.

	Background diffusion coefficient k_b	Initial strength of the THC	Critical E-P flux (at 55^oN) for the SST gradient collapse	THC for the critical E-P flux
1	$0.05x10^{-4}$ m^2/s	3-4Sv	0.8 m/year	has collapsed
2	$0.5x10^{-4}$ m^2/s	17Sv	1.1 m/year	is collapsing
3	$5x10^{-4}$ m^2/s	42Sv	2.5m/year	persists

Table 1. The main results of the freshening experiments for different values of the ocean vertical diffusion. While particular numbers can change from one model to another the main tendencies should persist.

Note that in the ocean model we use the full coefficient of vertical diffusion k is the sum of three components: background vertical diffusion k_b, the part dependent on the Richardson number (negligible except in the narrow equatorial region), and the part related to wind mixing (negligible below the first model layer). Except in the aforementioned few regions, the difference between k and k_b is small, so that we use the terms “background vertical diffusion” and “vertical diffusion” interexchangeably.

Another important factor that determines the ocean response to a freshening is the spatial structure of the forcing, that is, the location of areas with excess / deficit precipitation. Figure 6 shows some of the results of the experiments with an idealized, cosine-like structure of the anomalous freshwater forcing which has a maximum in precipitation at about 25^{o}N/S, and two minima, at the equator and at 45^{o}N/S. This shape, perhaps somewhat artificial, is chosen to impose surface freshening (negative E-P flux) over the regions of subduction associated with the wind-driven circulation, but to maintain excessive evaporation (positive E-P flux) over the regions of sinking associated with the thermohaline circulation. As one could expect, this experiment results in the collapse of the SST gradient along the equator (fig. 6, case 3), but the thermohaline circulation is affected only slightly as seen from summary plots in fig. 7. The mechanism of the equatorial SST gradient collapse is again related to the reduction of the meridional density gradient (between the subtropical subduction regions at about 30^o of latitude and the equator) and the deepening of the tropical thermocline, see fig. 7b. While the E-P anomaly used in these experiments is very idealized, such a result underlines the importance of where the freshwater forcing is applied.

Thus, an important finding of these experiments is that the collapse of the THC and that of the equatorial thermocline can occur independently. A simple explanation for this result is that the meridional density gradients controlling the THC and the wind-driven circulation are two different parameters that can vary independently from each other. For example, the forcing used in fig. 6 , does not change the meridional density gradient between 0^o and 60^{o}N where the sinking of the THC water occurs, but reduces the density gradient between the equator and the regions where water is subducted into the wind-driven circulation (at 30^o of latitude). In the experiments with different diffusion coefficients, the separation of the two gradients is more intricate, but the same arguments still apply.

One important note is in order: The results of perturbation experiments presented here correspond to 100 years of calculations (which makes them relevant to the climate change problem). Longer calculations show that in some cases, especially those near critical or transitional regimes, a steady state cannot be reached, but instead the system undergoes slow oscillations involving both the THC and the equatorial SST gradient. The oscillation period then ranges from centuries to millennia.

4. Results: Anomalous warming of high latitudes.

Numerous calculations with coupled GCM show a pronounced warming of the northern high latitudes in greenhouse-warming simulations (e.g. Stainforth *et al* 2005). To address this issue, in a separate series of experiments, we look at the effect of anomalous heating of the surface waters in high latitudes. An important prerequisite of these experiments is a superimposed weak E-P flux added when calculating the initial conditions. This additional "background" forcing, which corresponds to a freshening of surface water in high latitudes, remains constant from one experiment to the next (fig. 8b). The variable parameter in these perturbation experiments is the restoring temperature T*, chosen to induce a heating anomaly (with respect to the control case, i.e. experiment 1). The temperature anomaly is maximum at the northern and southern boundaries of the basin and tapers off at about 40^o of latitude (fig. 8a).

Since both freshening and warming of surface waters have similar effects on the density of sea water, one can expect that the results would be, to some degree, analogous to the experiments of the previous section. Indeed, our calculations show (again after 100 year of integrations) that when warmer high-latitude temperatures are imposed, the equatorial

SST gradient weakens (fig. 8c). This temperature gradient all but collapses for the warmest T* we imposed, and so does too the thermohaline circulation (not shown here).

The value of the maximum temperature anomaly we used in these experiments is 8^{o}C (or about 4^{o}C averaged over the area of the anomaly), which is relatively high. These values, however, would be smaller, had we chosen a somewhat stronger freshwater forcing. This and several other experiments suggest that the most efficient way to collapse both the THC and the equatorial SST gradient is to use a certain combination of anomalous surface freshening and warming. The equation of state of sea water, written as eq. (2), suggests that the effect of 1 psu (the unit of salinity) on density of sea water is equivalent to the effect of 3-5^{o}C, depending on mean temperature (see Fedorov *et al* 2004). Thus, one could vary either the freshwater or thermal forcing accordingly with the similar effects. Note, however, that if the freshwater forcing is completely absent, the meridional temperature gradient must vanish as well, for the THC and the equatorial SST gradient to collapse.

5. Conclusion

The effects of a freshening of the surface waters of the ocean in high latitudes have thus far been studied in connection with the thermohaline circulation, and separately in connection with the wind-driven circulation. In those modeling studies the focus has been on one or the other component of the oceanic circulation, because of the choice of certain model parameters. In the present study we have integrated the two approaches. This paper focuses on mechanisms that interfere effectively with the oceanic heat budget, thereby changing the thermal structure in the tropics, and thus influencing tropical sea surface temperatures and the global climate. The time-scale on which this can happen is that of the wind-driven circulation, a few decades. Hence these mechanisms amount to overlooked processes for rapid or abrupt climate changes.

Our experiments show that the freshening can induce either the collapse (weakening) of the THC with expected climate consequences, or the collapse (weakening) of the equatorial SST gradient leading to warm tropical conditions, or both. The factors that affect the strength of the necessary forcing include ocean diffusion, the relative magnitude of the freshening and warming anomalies (as they affect the oceanic heat transport) and the spatial structure of the forcing. Experiments with coupled General Circulation Models show some agreement with the results of this study. In particular, coupled simulations on the effect of high-latitude freshwater forcing demonstrate that, with a weakening of the THC, the Inter-tropical Convergence Zone (ITCZ) and the maximum of tropical precipitation shift southward in the Atlantic ocean (e.g. Vellinga and Wood 2002, Zhang and Delworth 2005). This would be consistent with the warming of the tropical SST in the eastern equatorial Atlantic controlled by changes in the wind-driven circulation. These important questions, as well as more realistic calculations, remain beyond the limits of the present study.

Several other issues remain unresolved. As the results of the present study stand now, they are applicable to the Atlantic ocean. The results of Fedorov *et al* 2004 are relevant to the Pacific. Both studies make the assumption of a closed basin. How restrictive is this? Can a freshening in the Atlantic result in the "permanent El Niño" in the Pacific? A tendency towards permanent El Niño is present, albeit weak, when the northern Atlantic is freshened in the recent study by means of a coupled climate model by Zhang and Delworth (2005), but the signal is more pronounced in the tropical Atlantic. In Dong and Sutton (2002) a temporary freshening of the northern Atlantic induces El Niño-like conditions in the equatorial the Pacific after 7 years. These authors suggested an atmospheric "bridge" as a possible explanation of the observed changes in the Pacific Ocean. An alternative explanation is that interfering with global oceanic heat transport leads to tropical changes in both basins (in other words, the surface heat fluxes in Figure 1 do not need to be balanced in each single basin separately). Today the Pacific gains heat that is exported to, and is lost in the Atlantic or the Southern Ocean. The connection between the oceans is through the Antarctic Circumpolar Current (Toggweiler and

Samuels 1995, Gnanadesikan 1999). Hence a freshening of the Atlantic that affects its heat budget could influence the Pacific. Recent results of Seidov and Haupt (2005) suggest that a differential inter-hemispheric freshening can be another important factor for the stability of the thermohaline circulation.

Appendix: Analytic expressions used in numerical calculations.

The wind stress used in this study is given by the following expression (Bryan 1987):

$$\tau^x = 0.8\,(-\sin(6|\phi|) - 1) + 0.5(\tanh(5\pi - 10|\phi|) + \tanh(10|\phi|)) \qquad \text{(A.1)}$$

where ϕ is latitude. The shape of the wind is symmetric with respect to the equator (see fig. 9).

The small E-P flux added to the forcing to induce preferential sinking in the Northern hemisphere is given by the following expression

$$F = 0.2*(\phi/\phi_{max}) \qquad \text{(A.2)}$$

where $\phi_{max} = 60^o$ and flux F is measured in m/year.

For the experiments presented in figs. 2-5 and 8, the shape of the E-P flux is calculated as

$$F = W_0 \cdot (1 - 2\,|\phi| / \phi_{max}) / cos(\phi) \qquad \text{(A.3)}$$

where $|\phi| \leq \phi_{max}$.

For the experiments presented in figs. 6 and 7, the shape of the E-P flux is calculated as

$$F = W_0 \cdot cos(2.5\pi\phi/ \phi_{max}) \qquad \text{(A.4)}$$

W_0 is a constant varied from one experiment to the next.

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References

Alley, R. and Coauthors, 2003: Abrupt Climate Change. *Science,* **299**, 2005-2010.

Boccaletti, G., R.C. Pacanowski, S.G. Philander and A.V. Fedorov. 2004: The thermal structure of the upper ocean. *J.Phys.Oceanogr.* **34**, 888-902.

Bryan, F., 1987: Parameter sensitivity of primitive equation ocean general circulation models. *J. Phys. Ocean.,* **17,** 970–985.

Curry, R. Dickson, B. and Yashayaev, I.: A change in the freshwater balance of the Atlantic Ocean over the past four decades. *Nature* **426**, 826 - 829, 2003.

Dickson B., Yashayaev I, Meincke J. *et al*. Rapid freshening of the deep North Atlantic Ocean over the past four decades. *Nature* **416**, 832-837, 2002.

Dong B.-W. and R.T. Sutton: Adjustment of the coupled ocean atmosphere system to a sudden change in the Thermohaline Circulation. *Geophys. Res. Letters* **29**, 2002.

Gnanadesikan, A., 1999: A simple predictive model for the structure of the oceanic pycnocline. *Science,* **283**, 2077-2079.

Fedorov, A. V., R. C. Pacanowski, S. G. Philander, and G. Boccaletti, 2004: The effect of salinity on the wind-driven circulation and the thermal structure of the upper ocean. *J. Phys. Oceanogr.* **34**, 1949–1966.

Fedorov, A. V. *et al* 2006: The Pliocene Paradox (Mechanisms for a permanent El Niño) *Science* 312, 1437-1443.

Harper, S., 2000: Thermocline ventilation and pathways of tropical - subtropical water mass exchange. *Tellus,* **52A**, 330-345.

Huang, R.X., 1988: Solutions of the ideal fluid thermocline with continuous stratification. *J. Phys. Oceanogr.,* **16**, 39-59.

Levitus, S. and T.P. Boyer, 1994c: World Ocean Atlas 1994 Volume 4: Temperature. NOAA Atlas NESDIS 4. U.S. Department of Commerce, Washington, D.C. 117 pp.

Luyten, J.R. and H. Stommel, 1986: Gyres driven by combined wind and buoyancy flux. *J. Phys. Oceanogr.* **16**, 1551-1560.

Luyten, J.R., J. Pedlosky and H.M. Stommel, 1983: The ventilated thermocline *J. Phys. Oceanogr.,* **13**, 292-309.

H. L. Johnson and D. P. Marshall, 2002: A theory for the surface Atlantic response to thermohaline variability. *J. Phys. Oceanogr.*, **32**, 1121-1132.

Knutson T.R., and S. Manabe: Time-mean response over the tropical Pacific to increased CO_2 in a coupled ocean-atmosphere model. *J.Climate* **8**, 2181-2199, 1995.

Manabe, S. and R.J. Stouffer, 2000: Study of abrupt climate change by a coupled ocean-atmosphere model. *Quaternary Science Reviews,* **19**, 285-299.

Manabe, S., and R.J. Stouffer, 1995: Simulation of abrupt climate change induced by freshwater input to the North Atlantic Ocean. *Nature,* **378**, 165-167.

Park, Y.-G., and K. Bryan, 2000: Comparison of thermally driven circulations from a depth-coordinate model and an isopycnal layer model. Part I: Scaling-law sensitivity to vertical diffusivity. *J. Phys. Oceanogr.,* **30,** 590–605.

Pedlosky, J., 1996: Ocean Circulation Theory. 453pp, Springer Verlag, Heidelberg.

Rahmstorf, S., 1995: Bifurcation of the Atlantic thermohaline circulation in response to changes in the hydrological cycle. *Nature,* **378**, 145-149.

Stainforth, D.A. and co-authors: Uncertainty in predictions of the climate response to rising levels of greenhouse gases. *Nature* **433**, 2005

Seidov, D. and B.J. Haupt, 2005: How to run a minimalist's global ocean conveyor *Geoph. Research Lett.,* 32, doi:10.1029/2005GL022559.

Seidov, D. and B.J. Haupt, 2003: On sensitivity of ocean circulation to sea surface salinity, *Global and Planetary Change,* **36**, 99-116.

Seidov, D. and B.J. Haupt, 2002: On the role of inter-basin surface salinity contrasts in global ocean circulation. *Geoph. Research Lett.,* **29**, doi: 10.1029/2002GL014813.

Stocker, T.F. and A. Schmittner, 1997: Influence of CO2 emission rates on the stability of the thermohaline circulation. *Nature,* **288**, 862-865.

deSzoeke, R., 1995: A model of wind and buoyancy-driven ocean circulation. *J. Phys. Oceanogr.,* **25,** 918–941.

Toggweiler, J.R., and B. Samuels, 1998: On the ocean's large-scale circulation near the limit of no vertical mixing. *J. Phys. Oceanogr.,* **28**, 1832-1852.

Veronis, G., 1988: Circulation driven by winds and surface cooling. *J. Phys. Oceanogr.* 18, 1920-1932

Veronis, G., 1976. Model of world ocean circulation: III. Thermally and wind-driven. *J.Marine Research* **36**, 1-44.

Vellinga, M., and R. A. Wood, 2002: Global climatic impacts of a collapse of the Atlantic thermohaline circulation. *Climatic Change* **54**, 251-267.

Zhang, R., and T. L. Delworth, 2005: Simulated tropical response to a substantial weakening of the Atlantic thermohaline circulation. *J. of Climate* **18**, 1853-1860.

Zhang, J., R.W. Schmitt and R.X. Huang, 1999: The relative influence of diapycnal mixing and hydrologic forcing on the stability of the thermohaline circulation. *J. Phys. Oceanogr.,* **29**, 1096-1108.

Figures:

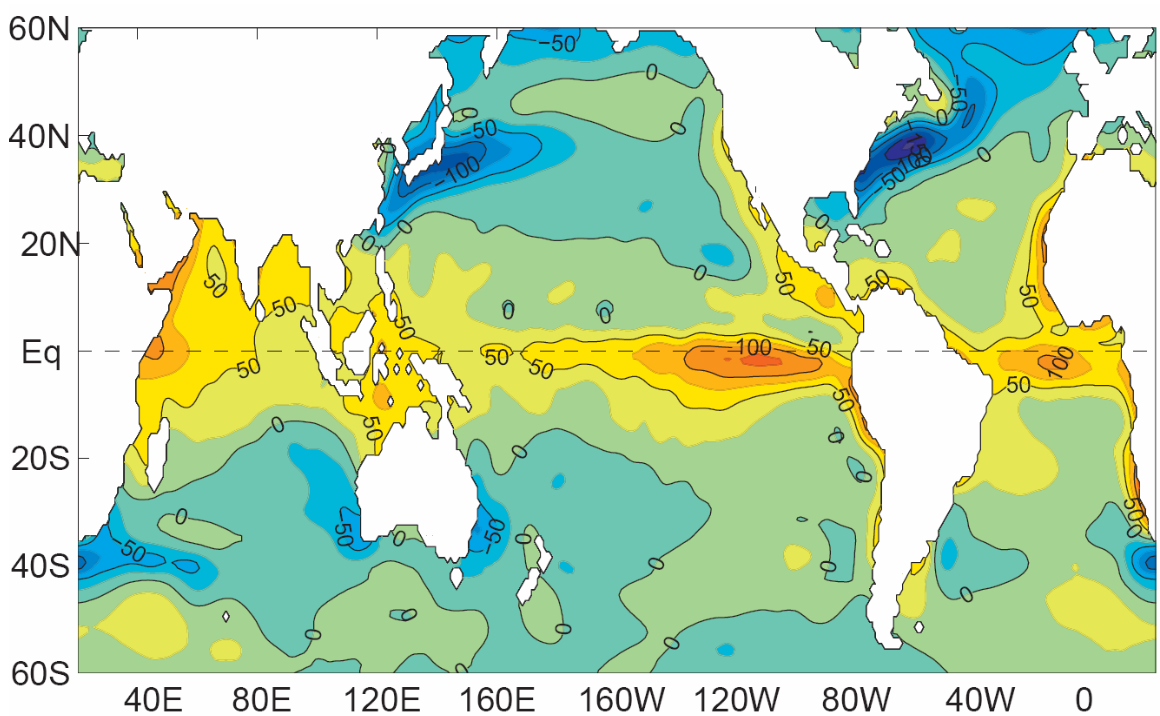

Figure 1. Annual mean net heat flux across the ocean surface in watts/m^2 (after Da Silva *et al*., 1994). The heat gain in the upwelling zones of low latitudes is roughly equal to the heat loss in high latitudes. In a steady state an increased heat loss in high latitudes implies a larger poleward heat transport from the tropics. (The ocean heat transport through a given latitude is an integral of the surface flux from the equator to this latitude.) Exactly how the ocean gains heat depends on its vertical diffusivity. If this parameter is small, then the thermocline is sharp and has to be shallow for the ocean to gain a large amount of heat. If the diffusivity is large then the ocean gains heat by diffusing it downwards into the deep ocean.

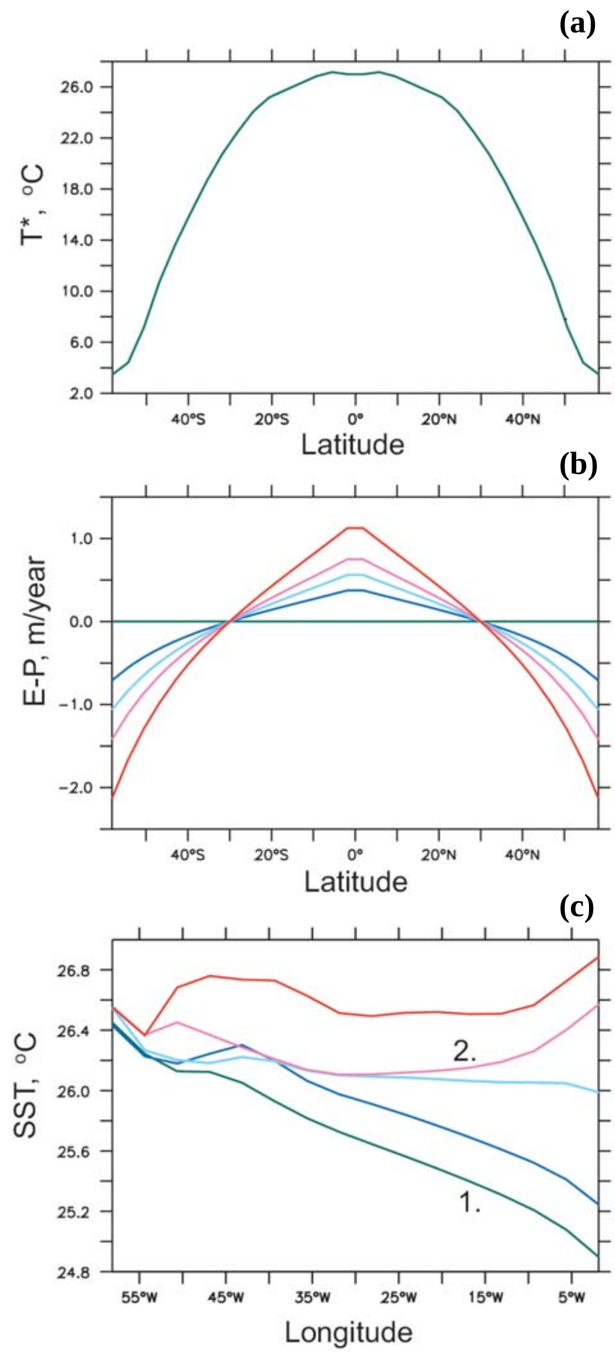

Figure 2. The forcing, and the ocean response in the equatorial region, for different values of the freshwater flux in high latitudes. Intermediate values of background diffusion, k_b= 0.5x10^{-4}m^2/s, are used. The color of each line in (c) matches the respective color in (b). Case 1: the anomalous freshwater forcing is zero. Case 2: the SST gradient along the equator has collapsed.

(a) The meridional structure of the restoring temperature T*.
(b) The meridional structure of the anomalous E-P flux (in m/year). Negative values indicate an excess of precipitation over evaporation.
(c) The SST along the equator after 100 years of integrations.

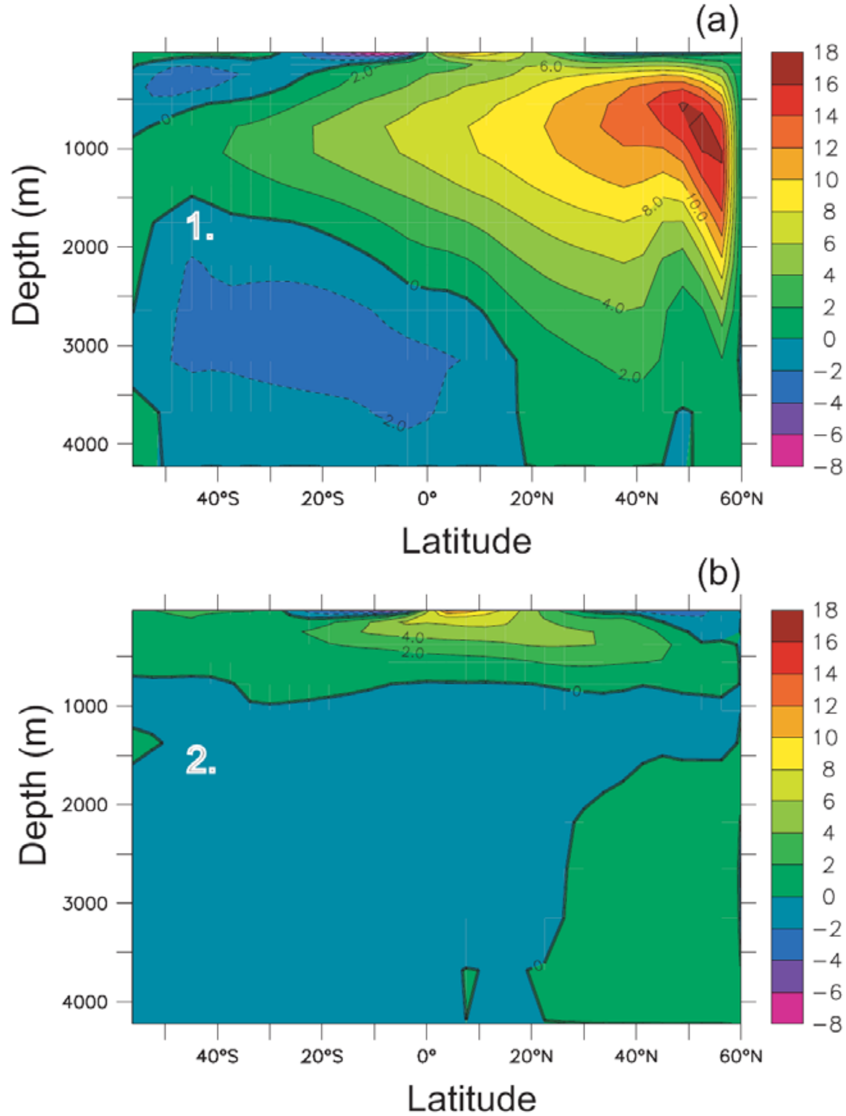

Figure 3. The meridional overturning circulation (in Sv) for experiments 1 and 2 after 100 years of calculations. Notice a deep overturning associated with the THC (top panel) and a shallow overturning (upper 200m) associated with the wind-driven component of the circulation (both panels). The THC has collapsed in the bottom panel.

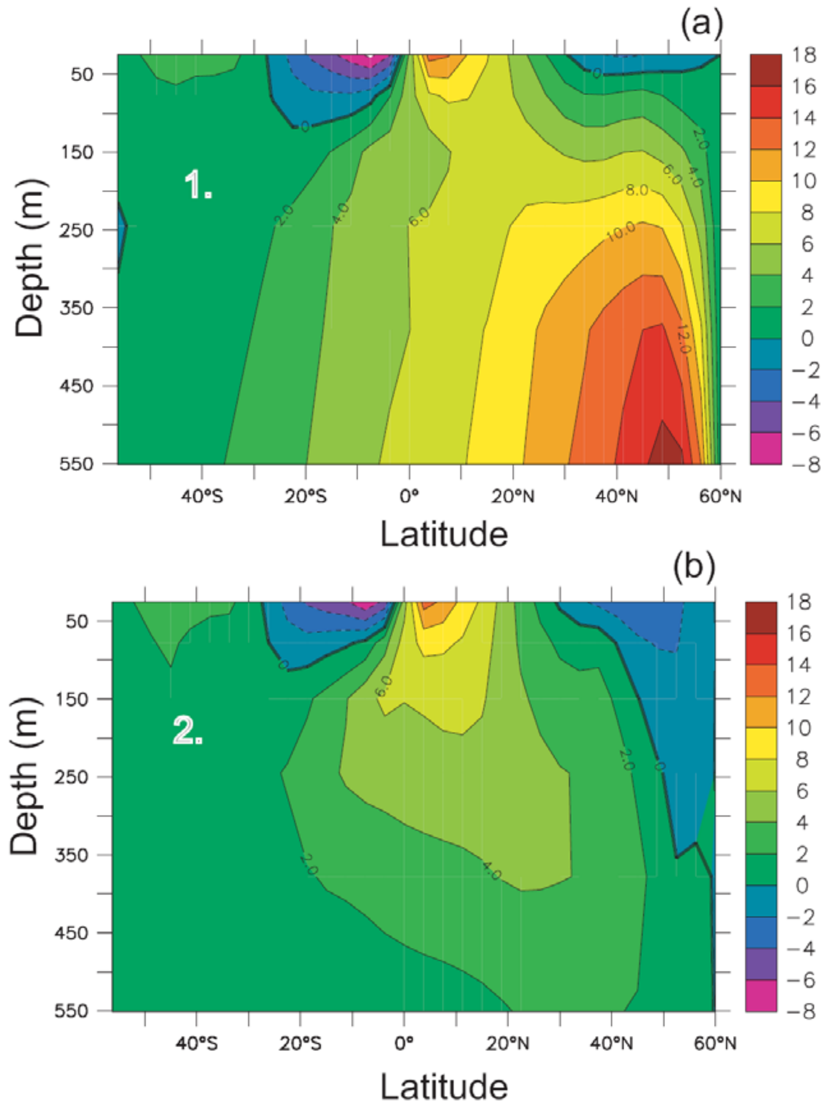

Figure 4. A detailed structure of the meridional overturning (in Sv) for the upper 500m of the ocean, for the cases shown in figure 3. Notice the shallow wind-driven circulation (the upper 200m) and the upper branch of the deep overturning associated with the THC (top panel). The shallow wind-driven cell effects a large part of the heat transport from the tropics. The THC has collapsed in the bottom panel.

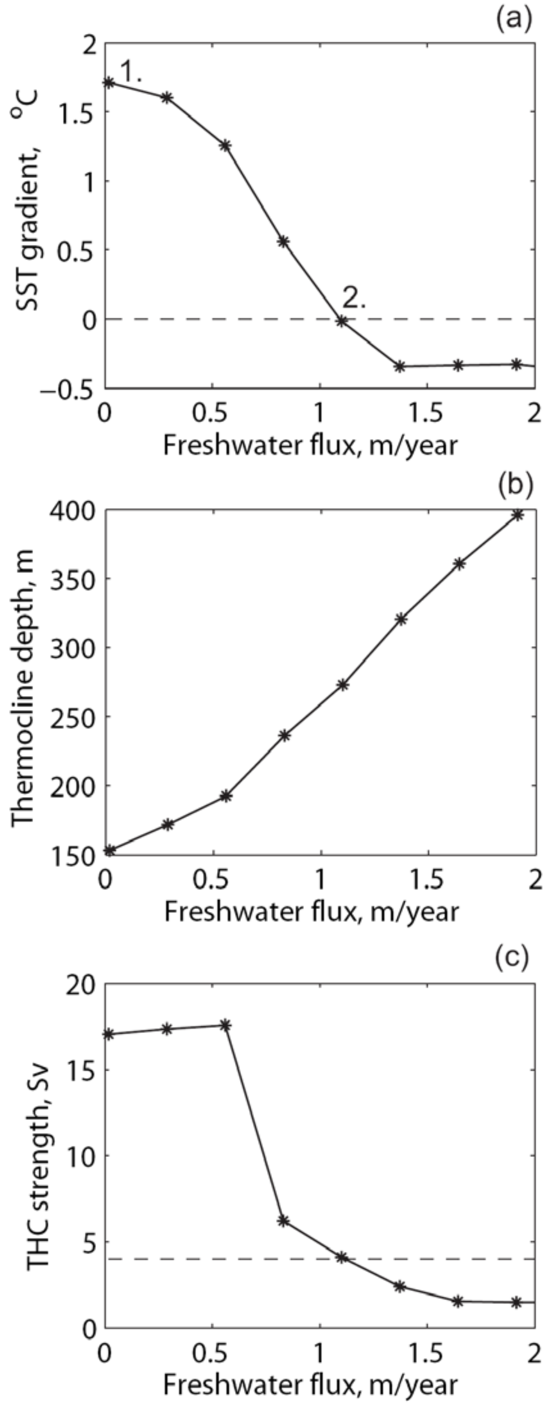

Figure 5. A summary plot for several perturbation experiments including those shown in figure 2 (cases 1 and 2 are specifically marked). The freshwater flux is measured at 55^{o}N.

(a) Reduction and reversal of the zonal SST gradient along the equator.
(b) Deepening of the tropical thermocline (thermocline depth is measured in the middle of the basin).
(c) Reduction and collapse in the strength of the THC (dashed line corresponds to 4Sv).

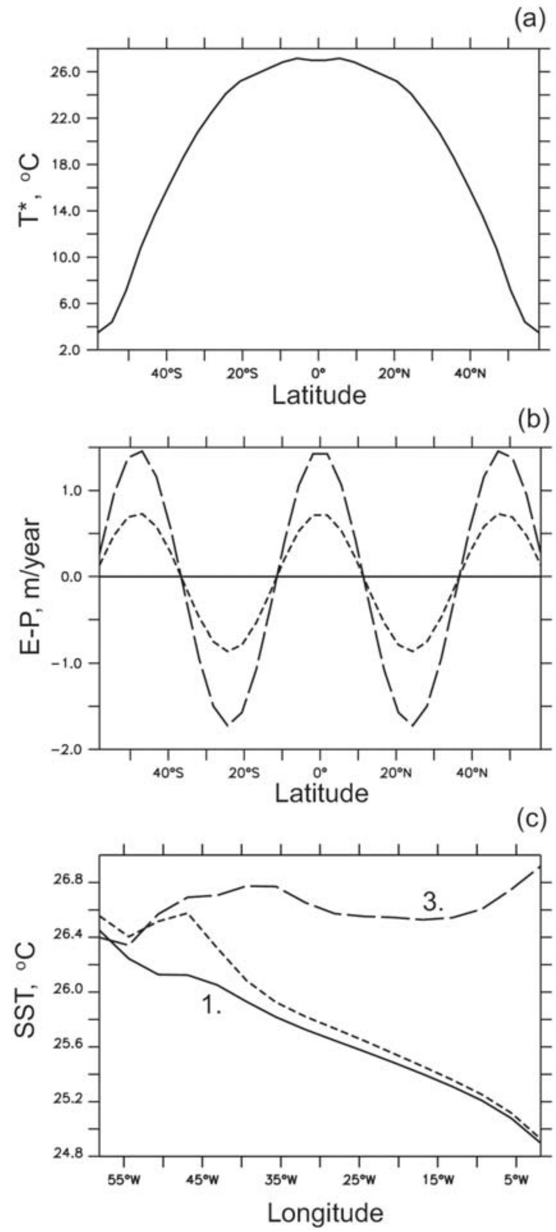

Figure 6. The forcing, and the ocean response in the equatorial region, for different amplitudes of the cosine-like E-P forcing. The structure of the E-P flux is given by equation (A.3) of the Appendix. The restoring temperature T* is the same as in fig. 2. Case 1: the anomalous freshwater forcing is zero. Case 3: the SST gradient along the equator has collapsed, but the THC is affected only slightly (see fig. 6B). For intermediate values of background diffusion, $k_b = 0.5 \times 10^{-4} m^2/s$:

(a) The meridional structure of the restoring temperature T*.
(b) The meridional structure of the anomalous E-P flux (in m/year). Negative values indicate an excess of precipitation over evaporation.
(c) The SST along the equator after 100 years of integrations. The style of each line in (b) matches the respective line style in (c).

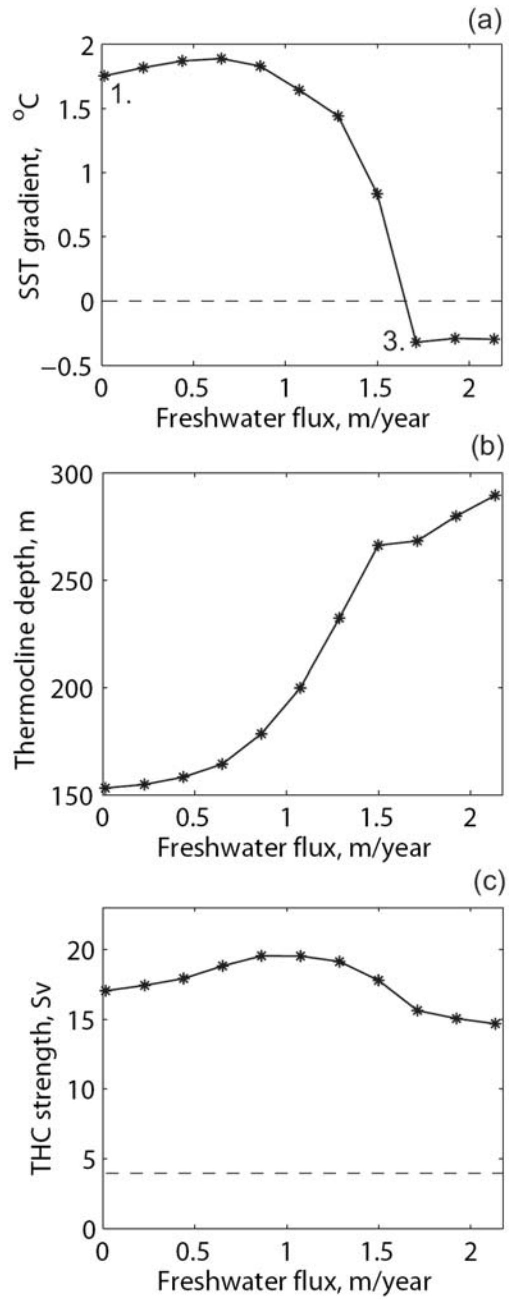

Figure 7. A summary plot for several perturbation experiments with a cosine-like forcing including those shown in figure 6 (cases 1 and 3 are specifically marked). The freshwater flux is measured at 25^{o}N.

(a) Reduction and reversal of the zonal SST gradient along the equator.
(b) Deepening of the tropical thermocline (thermocline depth is measured in the middle of the basin).
(c) Changes in the strength of the THC (dashed line corresponds to 4Sv).

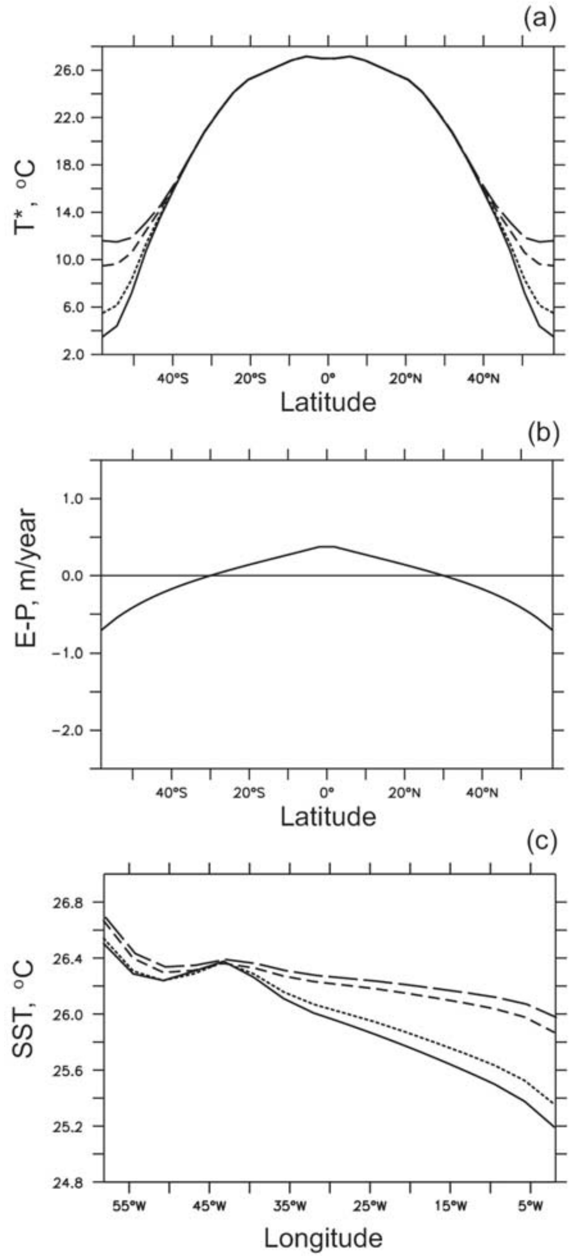

Figure 8. The forcing, and the ocean response in the equatorial region, for different values of the anomalous heating in high latitudes. For intermediate values of background diffusion $k_b=0.5x10^{-4}m^2/s$:

(a) The meridional structure of the restoring temperature T* in perturbation experiments.
(b) The meridional structure of the background E-P flux (in m/year). Negative values of E-P indicate an excess of precipitation over evaporation.
(c) The SST along the equator after 100 years of integrations. The style of each line in (c) matches the respective line style in (a).

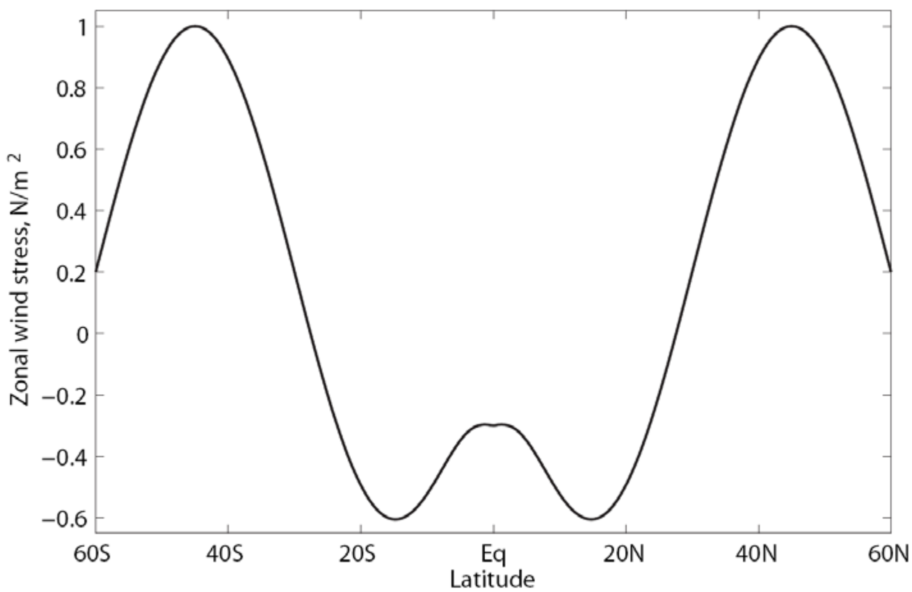

Figure 9. The analytic approximation to the zonally-averaged annual mean wind stress used in this study (after Bryan 1987).