

The Thermal Structure of the Upper Ocean

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(Manuscript received 19 November 2002, in final form 3 September 2003)

ABSTRACT

The salient feature of the oceanic thermal structure is a remarkably shallow thermocline, especially in the Tropics and subtropics. What factors determine its depth? Theories for the deep thermohaline circulation provide an answer that depends on oceanic diffusivity, but they deny the surface winds an explicit role. Theories for the shallow ventilated thermocline take into account the influence of the wind explicitly, but only if the thermal structure in the absence of any winds, the thermal structure along the eastern boundary, is given. To complete and marry the existing theories for the oceanic thermal structure, this paper invokes the constraint of a balanced heat budget for the ocean. The oceanic heat gain occurs primarily in the upwelling zones of the Tropics and subtropics and depends strongly on oceanic conditions, specifically the depth of the thermocline. The heat gain is large when the thermocline is shallow but is small when the thermocline is deep. The constraint of a balanced heat budget therefore implies that an increase in heat loss in high latitudes can result in a shoaling of the tropical thermocline; a decrease in heat loss can cause a deepening of the thermocline. Calculations with an idealized general circulation model of the ocean confirm these inferences. Arguments based on a balanced heat budget yield an expression for the depth of the thermocline in terms of parameters such as the imposed surface winds, the surface temperature gradient, and the oceanic diffusivity. These arguments in effect bridge the theories for the ventilated thermocline and the thermohaline circulation so that previous scaling arguments are recovered as special cases of a general result.

1. Introduction

The thermocline is so remarkably shallow in the Tropics and subtropics that the average temperature of the water column, even in the western equatorial Pacific where surface temperatures are at a maximum, is barely above freezing. The oceanic circulation that maintains this thermal structure has two main components, a shallow wind-driven circulation and a deep thermohaline circulation. Traditionally, theoretical models of the thermocline are classified according to their focus on one or the other of these components.

Theories that focus on the thermohaline circulation in a closed basin (e.g., Bryan 1987) assume that cold water, originating from high latitudes and flooding the abyss, rises uniformly everywhere and is heated by the

downward diffusion of heat. An expression for the depth of the thermocline can then be obtained by assuming geostrophic, hydrostatic, incompressible motion in accord with the balance

$$fu_z = -g\gamma T_y, fv_z = g\gamma T_x, \text{ and } \beta v = fw_z, \quad (1)$$

where u and v are the horizontal velocities, w is the vertical velocity, T is the temperature (we shall ignore salinity), g is the acceleration of gravity, γ is the coefficient of thermal expansion of water, f is the Coriolis parameter, and β is its meridional gradient. The thermodynamic equation is vertical advection diffusion:

$$wT_z = \kappa T_{zz}. \quad (2)$$

Equations (1) and (2) result in an estimate for the depth of the thermocline:

$$D = [\kappa f^2 L / (g\beta\gamma\Delta T)]^{1/3}, \quad (3)$$

where ΔT is the static stability of the ocean, L is the horizontal dimension, and D is the depth of the thermocline.

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The estimate in Eq. (3) may be relevant to a highly diffusive ocean, but reality appears to correspond to the limit of low diffusivity (Ledwell et al. 1993). Furthermore, Eq. (3) does not reflect the influence of the wind on the oceanic circulation and its thermal structure. Let us therefore turn our attention to the shallow, wind-driven circulation of the ventilated thermocline. It involves subduction into the thermocline in certain subtropical regions and the return of water to the surface layers in upwelling zones such as that of the eastern equatorial Pacific. Theories for thermal structure maintained by such a circulation are implicit in the work of Welander (1959) and were developed by Luyten et al. (1983) and Huang (1986). These are often referred to as adiabatic theories because of the fundamental hypothesis that the flow in the main thermocline is conservative; in these theories the thermal structure is determined by wind-driven processes.

Equations (1) are again appropriate for this circulation. However the flow is assumed to be conservative so that the thermodynamic equation is now

$$uT_x + vT_y + wT_z = 0. \quad (4)$$

Equations (1) and (4) can be solved analytically (Luyten et al. 1983; Huang 1986). At the upper boundary the vertical velocity is set to be the wind-forced Ekman pumping $w = w_{\text{Ek}}$. In the simplest case of a two-layer system, the depth of the thermocline is

$$D = \left(\frac{w_{\text{Ek}} f L^2}{g \gamma \Delta T} + D_e^2 \right)^{1/2}, \quad (5)$$

where D_e is an arbitrary constant representing the depth of the thermocline on the eastern boundary. Although the flow is assumed to be adiabatic [Eq. (4)], thermodynamic processes are implicit in the solution. Different values of the parameters will lead to different thermocline depths and therefore different oceanic heat content. Diabatic processes must be present to accommodate the conversion of water associated with different thermocline depths. Because these thermodynamic processes are not taken into account, the solution is not complete and D_e remains an arbitrary constant.

Attempts to represent both the wind-driven and the thermohaline circulations in one coherent picture have been made by Robinson and Stommel (1959) and by Salmon (1990). The culmination of those efforts leads to the proposal by Samelson and Vallis (1997) of a two-thermocline limit: the adiabatic theories are appropriate in the upper part of the thermocline, where ventilation is dominant, and the diffusive theories are appropriate in the lower part of the thermocline, where there is no ventilation.

Adding diffusion to a primarily adiabatic theory does not however resolve the issue of arbitrary background stratification [see, however, Tziperman (1986) and de Szoeke (1995) for specific cases]. Samelson and Vallis (1997) proposed that the eastern boundary thermocline

depth behaves as described by Pedlosky (1987), but this, as we shall see, is also a very special case, with possibly limited applicability to the real ocean. Therefore the thermocline problem is still very much open, despite these advances in our theoretical understanding.

A theory that accounts for the thermal structure of the ocean must explain not only the depth of the thermocline D , but also ΔT , the oceanic static stability, a critical parameter for both the diffusive estimates in Eq. (3) and the adiabatic one in Eq. (5). Lionello and Pedlosky (2000) showed how this ΔT controls the character of the solution in the adiabatic ventilated thermocline theory. Implicit in this parameter is recognition that a thermohaline process is maintaining the abyss at the appropriate temperature, but what is the relationship between the solution given in Eq. (5) and the thermohaline circulation implied by ΔT ? What determines ΔT ?

This paper attempts to integrate the approaches to the thermocline problem discussed thus far by invoking the constraint of a balanced heat budget. The focus of previous theories has been largely on satisfying the constraints imposed on the flow by Eqs. (1). The global thermodynamic behavior of the ocean circulation is at best implicit in such an approach. Explicit treatment of thermodynamic processes will allow us to construct a theory for the thermal structure of the ocean and the external parameters that control it. Far from being an alternative to previous theories this work purports to complement those efforts by explicitly considering the implications of an aspect of the global circulation thus far largely ignored.

The limitations of considering simplified models of ocean circulation such as those represented by Eqs. (1), (2), and (4) will be considered in the final section. In the next section we will present a brief analysis of the oceanic heat budget, followed by a statement of the hypothesis and an outline of the rest of the work.

2. Surface heat fluxes and the constraint of a balanced heat budget

Figure 1 shows the climatological surface annual mean heat flux.

Shortwave radiation and latent heat loss are the dominant terms in the balance (Oberhuber 1988; da Silva et al. 1994). Over most of the domain, these two terms are in approximate local balance so that over the course of a year the net flux across the ocean surface is zero. There are, however, notable exceptions to this local equilibrium as shown in Fig. 1. Large amounts of heat are lost on a yearly average in the neighborhood of the western boundary currents, where those currents separate from the coast. There the ocean loses large amounts of heat when cold dry winds blow from the continent over the warm waters carried by the currents, inducing a large latent heat loss to the atmosphere.

Heat gain is concentrated in the tropical bands of the Pacific and Atlantic Oceans in the eastern side of the

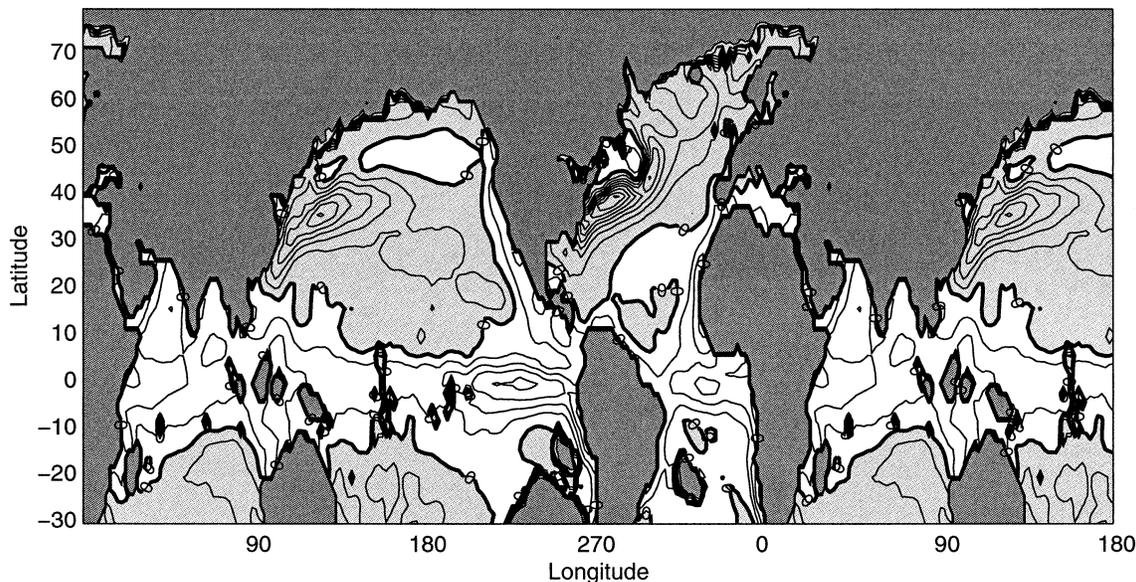


FIG. 1. Annual mean net surface heat flux (from Oberhuber 1988). Contour interval: 25 W m^{-2} .

basin along the equator and along the eastern coasts. These are upwelling zones, where low sea surface temperatures inhibit evaporation, so that latent heat loss to the atmosphere does not balance insolation. The ocean then gains heat as described, for example, in Bunker (1976).

The surface temperatures in those regions therefore control the amount of heat entering the ocean. The surface temperatures are in turn the result of the vertical thermal structure of the ocean: cold water upwells because the thermocline is sufficiently shallow to allow the entrainment of cold water into the mixed layer. Regions of upwelling are therefore special in that the vertical thermal structure is directly related to the heat budget. This is particularly evident during El Niño events when the interannual modulation of thermocline depth modifies the sea surface temperature distribution. The cold tongue disappears and promptly the heat gained by the ocean is reduced (Weare 1984; Philander and Hurlin 1988).

In steady state, the entrainment of cold water in the tropical and boundary mixed layer must be maintained by a subsurface flow fed by the subduction of cold, surface waters in higher latitudes (Wyrki 1981). Regions of upwelling, where the heat is gained, and mid-latitude regions, where the heat is lost, must therefore be connected if the ocean is in steady state.

Notice that, if the regions of heating and cooling were not balanced, the unbalanced heating would have to result in a change in heat content and therefore in a change in depth and structure of the thermocline. In the absence of oceanic heat transport, a heating of 50 W m^{-2} results in a deepening of the thermocline of about 40.0 m yr^{-1} (the temperature difference across the thermocline is assumed to be 10°C)! The maximum heating

in the Tropics exceeds 100 W m^{-2} , and yet the climatological position of the mean thermocline has hardly changed in the last 50 years (Harper 2000). This fact suggests that a very tight connection must exist between regions of heating and cooling, tight enough to maintain the heat content of the upper ocean as approximately stationary.

The vertical thermal structure of the ocean is directly coupled to the heating in the Tropics and along the boundaries, so the constraint of a balanced heat budget must constrain the depth of the thermocline. In the next sections we shall turn to a general circulation model of the ocean to investigate more carefully the relationship between a balanced heat budget and the vertical thermal structure of the ocean. The model is described in detail in section 3, and the simulations are described in section 4.

A theory to try to explain the behavior of the model is derived in section 5. A discussion of the theory is provided in section 6. Conclusions are presented in section 7.

3. The model

The model is very idealized and provides a conceptually simple setting in which to test the hypothesis. A small tropical basin is forced by uniform easterly winds and an idealized thermal boundary condition. The basin extends from 16°N to 16°S and is 40° wide and 5000 m deep. The choice of such a small basin allows a large number of cases to be integrated.

The wind stress is 0.05 Pa , and, because it is constant, the complication of a full gyre circulation and boundary current is largely avoided. Winds force upwelling within a Rossby radius of the equator and drive surface Ekman flow toward the southern and northern boundaries. Be-

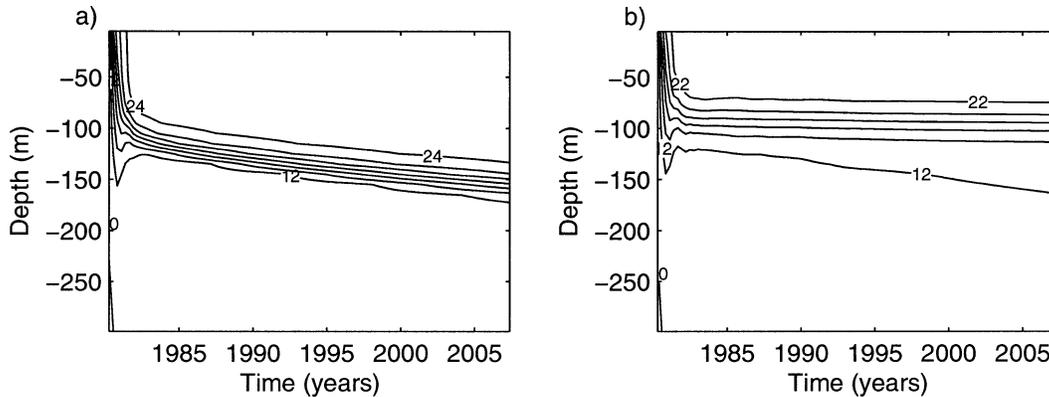


FIG. 2. Time evolution from 10°C isothermal conditions for two different cases (the time evolution is for the thermal structure at one point at the equator). (a) Thermal structure at 20°E on the equator for the case with uniform restoring temperature at 25°C. (b) As in (a) but for the case in which the restoring temperature tapers down to 10°C poleward of 12°N and 12°S.

cause no vorticity is imparted by the winds, convergence of Ekman flow, due to the sphericity of the earth, produces downwelling off the equator and an equatorward flow that balances the Ekman flow in the mixed layer. The winds maintain an equatorial undercurrent.

The thermal boundary condition is provided by restoring surface temperatures to an imposed temperature distribution (Haney 1971). This formulation is the simplest possible interactive atmosphere, one with infinite heat capacity, where heat fluxes respond to changes in sea surface temperatures as expected from a linearized form of the surface heat budget. The heat flux is chosen to be

$$Q = \alpha(T^* - T), \quad (6)$$

where $\alpha = 50 \text{ W (K m}^2\text{)}^{-1}$, which restores surface temperature T to a specified T^* .

The model used is the Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model 4 (MOM4). Horizontal resolution is specified as $\frac{1}{2}^\circ$ everywhere, with 32 levels in the vertical direction. The vertical resolution is 10 m in the upper 200 m.

The initial temperature is a uniform 10°C, unless otherwise specified. Vertical mixing is the Pacanowski and Philander (1981) scheme solved using a maximum vertical mixing coefficient of $50 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ and a background diffusivity of $0.01 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$, unless otherwise specified. Constant lateral mixing coefficients are $A_m = 2.0 \times 10^3 \text{ m}^2 \text{ s}^{-1}$ and $A_h = 1.0 \times 10^3 \text{ m}^2 \text{ s}^{-1}$ for momentum and heat, respectively.

All statements are made for quasi-steady-state circulations. Ideally all solutions should be run to equilibrium. However, even in the small basin used for these experiments the time scale for equilibrium is very long and computationally expensive, especially for simulations with low diffusivity.

Our focus is on the sensitivity of the upper-ocean thermal structure to surface forcing. We shall therefore take the working definition of steady state as a state in

which the simulation is approximately in balance with the wind forcing and the integral of the heat budget is small compared to the absolute values of the maxima and minima. In such experiments our model is integrated for 30 years. This approach exploits the dramatic time scale difference between processes that affect abyssal temperatures and the processes that are responsible for determining the upper thermocline.

This can be seen in the results from a simple experiment, which demonstrates that the heat budget, as a global constraint, is a critical factor in determining the stratification of the ocean. Assume first that temperatures at the surface are restored everywhere toward $T^* = 25^\circ\text{C}$. In due course the equilibrium solution will necessarily be an isothermal 25°C as the model can only warm up. The final solution therefore will have no thermocline and the circulation will penetrate to the bottom. In Fig. 2a this is seen to happen on a time scale far longer than the remarkably short time scale for the wind-driven circulation to come into adjustment and for a thermocline to appear. A thermocline is established very rapidly, within the first three years, and then the thermocline continues to deepen steadily on a much longer time scale. If run for a sufficiently long time, the thermocline will continue to deepen and expand until the entire basin asymptotically reaches a uniform temperature of 25°C .

Now consider a case in which the restoring temperatures at high latitudes in the model (poleward of 12°) are linearly tapered down to 10°C at the northern and southern boundaries. The final solution, which we will analyze in more detail later, is shown in Fig. 3. A clear thermocline is present and the circulation is largely confined to the upper ocean. In this second case (Fig. 2b) both warming and cooling are present, and the thermocline is locked in place immediately after the first phase of the adjustment. The basin continues to warm up as heat is transferred to the abyss, and the process continues until the boundary layer between the venti-

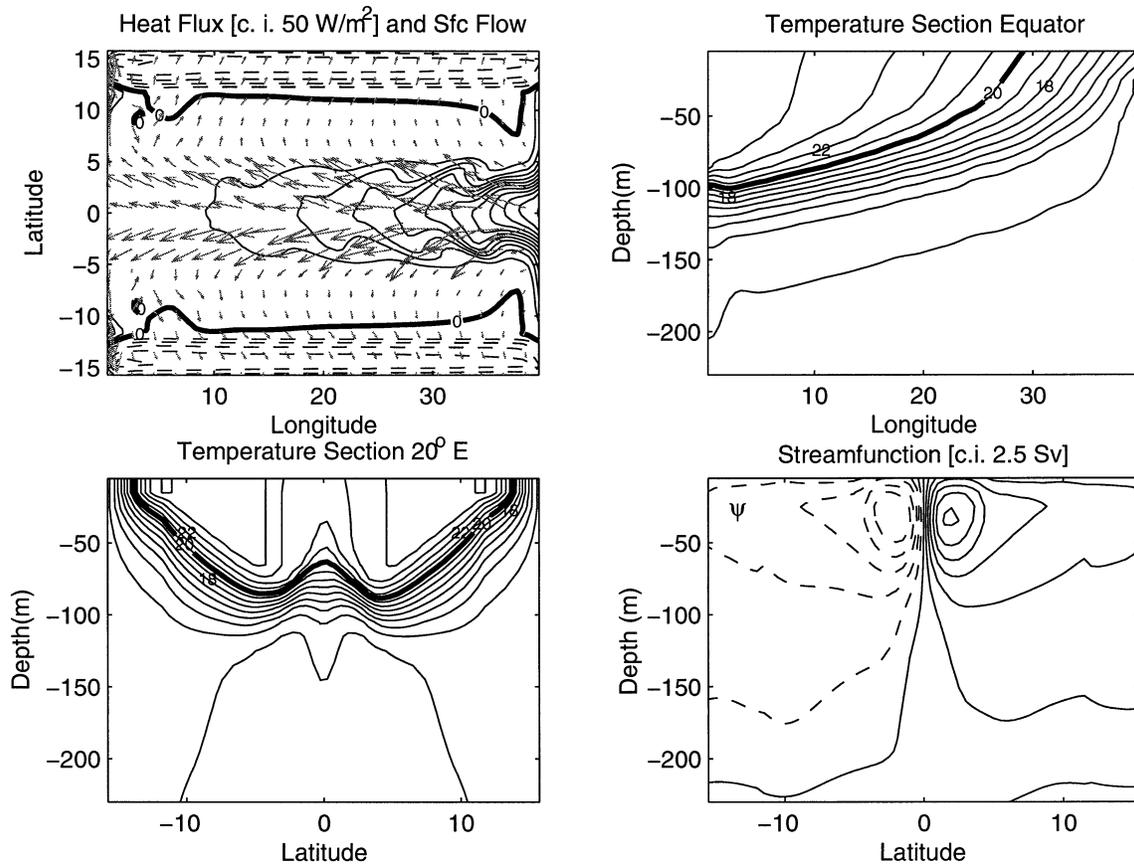


FIG. 3. Mean structure of the steady-state solution for the default case.

lated thermocline and the abyss satisfies a local advective–diffusive balance. However, the ventilated thermocline is essentially unaffected by this process, and its final structure (after 500 years) is not significantly different from the one obtained immediately after the first few years of the adjustment.

The only difference between the two cases is in the thermal boundary condition at high latitudes: in one case cooling is allowed whereas in the other it is not. Both models are subject to the same mechanical forcing, but the solutions are dramatically different, even though both models start from the same initial condition. We will not dwell on the details of the adjustment, as they will be analyzed in detail by Boccaletti (2004, manuscript submitted to *Dyn. Atmos. Oceans*). The important point for the purposes of this study is that a simple change in the thermal boundary condition has a profound effect on the structure of the thermocline, and any theory that purports to explain the structure of the thermocline should explicitly take these effects into account.

4. Numerical experiments

In this section we investigate the conditions under which a steady-state thermal structure of the model

ocean is modified and what external parameters control the stratification.

a. The mean state of the model

The steady-state circulation associated with the adjustment of Fig. 2b is shown in Fig. 3. The constant easterly wind drives a poleward Ekman transport, which encounters a temperature gradient poleward of 12°N and 12°S . This results in cooling of surface waters and downwelling at the boundary. The flow then returns at depth toward the equator, where it is first fed into the undercurrent and then upwelled to the surface in a large cold tongue. The surface heat flux reflects such a circulation, with a more or less zonally uniform cooling in high latitudes and a warming at the equator that matches the surface structure of the cold tongue. The cold tongue itself is the result of outcropping of the equatorial thermocline, the tilt of which is approximately set by the winds.

b. Sensitivity to the restoring temperature

To modify the thermal structure of the solution shown in Fig. 3, the heat budget at the surface was changed to increase heat loss in high latitudes. All else being

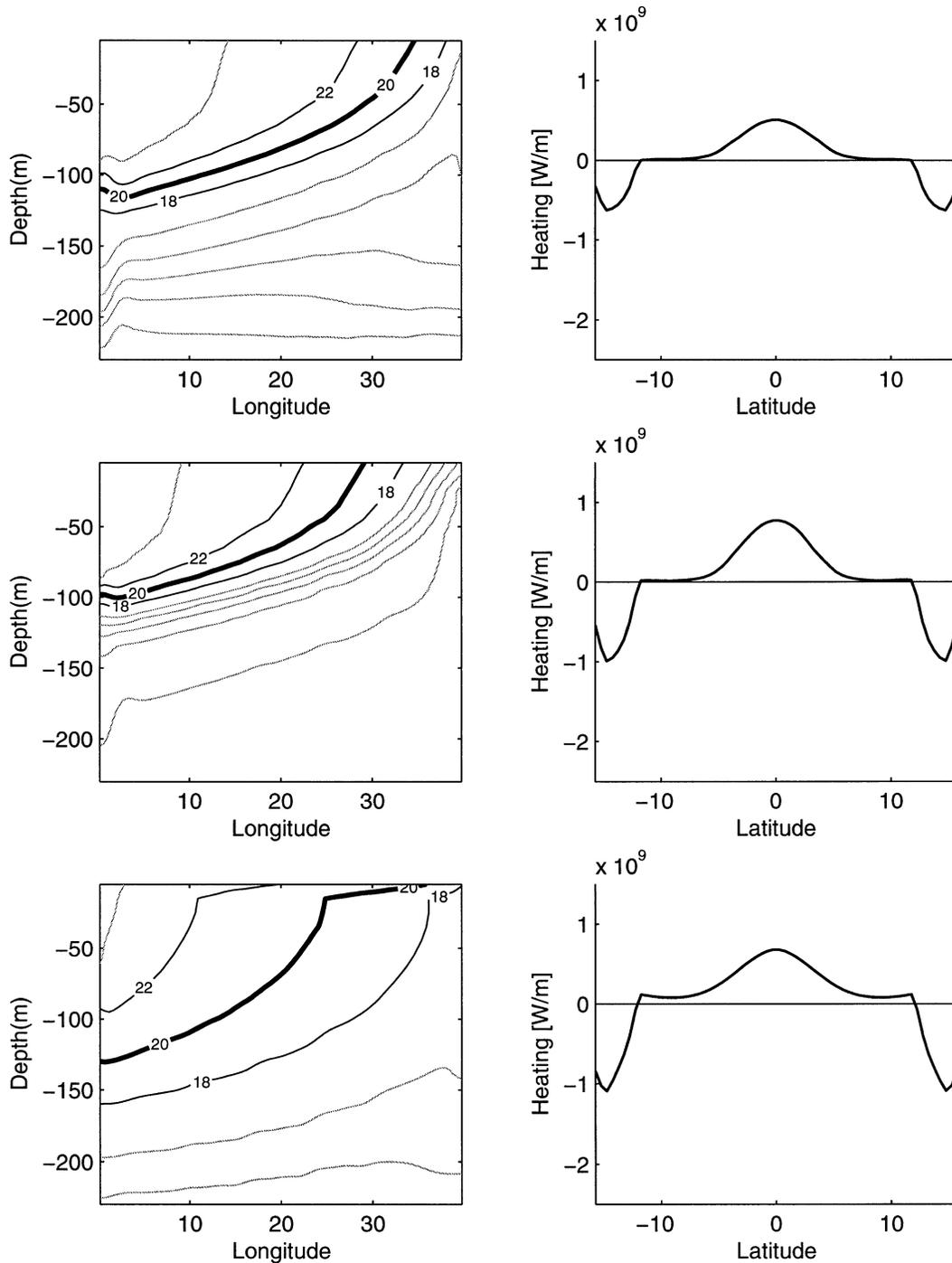


FIG. 4. (right) Zonally integrated heat fluxes, and (left) corresponding thermal structure at the equator. (top) Restoring the high-latitude temperature toward 15°C, (middle) restoring the high-latitude temperature toward 10°C, and (bottom) same as (middle) but for high diffusivity $\kappa = 10^{-4} \text{ m}^2 \text{ s}^{-1}$.

fixed, the result is a shoaling of the thermocline so that heating in the Tropics increases.

Two such cases are shown in Fig. 4, where the high-latitude restoring temperature is set to 15°C (top) and 10°C (middle). The high-latitude cooling of the ocean

is increased by increasing the temperature gradient that the Ekman flow encounters.

The increase in high-latitude cooling provokes a response in the tropical regions of upwelling. The equatorial thermocline shoals to expose more cold water,

thereby increasing the heating to minimize the imbalance in the heat budget. It should be emphasized that neither winds nor restoring temperatures have been changed in the Tropics, and yet the tropical ocean undergoes change.

Because of the colder high-latitude temperatures the shoaling of the thermocline is also accompanied by an increase in the static stability of the ocean (ΔT across the thermocline). Therefore, this behavior is also consistent with the notion that the depth of the thermocline represents the depth of penetration of the wind forcing and is, therefore, inversely proportional to the strength of the stratification, as seen for instance in the wind-driven scaling. However, this would not be a satisfying explanation because, as the following experiment suggests, the balance cannot be merely mechanical. The same change in high-latitude cooling can be produced in a set of simulations with higher diffusivity ($\kappa \sim 10^{-4} \text{ m}^2 \text{ s}^{-1}$ rather than $0.01 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$). For this case, we give solutions from an experiment in which the high-latitude restoring temperature is 10°C , as shown in Fig. 4 (bottom).

Higher diffusivity, for the same boundary conditions, results in a remarkably different static stability and stratification. In this case, the thermocline is more diffuse and deeper. Heat loss in high latitudes is comparable in both, but in the high-diffusion case it is balanced partly by heating in the cold tongue and partly by heating over a much larger area associated with diffusion across the thermocline into the abyss. In contrast, in the low-diffusion case, most of the heating occurred by exposing water that is colder than the restoring temperature to the surface, in the region of the cold tongue.

These results suggest that the partitioning of heat transport between the upper wind-driven circulation and the deeper thermohaline process depends strongly on diffusion. Even a theory focusing on the wind-driven circulation should include diffusive processes explicitly, because they are critical at least in the high-diffusion limit.

c. Sensitivity to wind strength

If the mechanical forcing is increased, a number of effects are expected. First, a stronger Ekman flow will transport more warm water poleward, resulting in a stronger cooling for the same restoring temperature. Also upwelling will be stronger, as an increase in wind strength will result in stronger equatorward subsurface return flow.

The sensitivity to the strength of the winds is shown in Fig. 5 for two different easterly wind stresses: 0.0125 and 0.05 Pa.

Complicating the interpretation of the response is the fact that winds not only change the strength of the equatorial upwelling but also the zonal gradient (tilt) in the equatorial thermocline. If the buoyancy budget were fixed, a strengthening of the winds would result in an

increased tilt of the thermocline and therefore a deepening of the thermocline to preserve the area of the cold tongue. In our experiments, however, an increase in tilt due to the strengthening of the winds corresponds to an increased heat loss, determined by a strengthened Ekman flow driven by those same winds. This in turn should result in a shoaling of the thermocline so that more heat can be gained in the Tropics.

The combination of these two effects in the experiment results in a deeper thermocline and a stronger heat transport. This is because the cooling effect due to the upwelling of cold water leads to a stronger heating than required by the heat loss so that the thermocline adjusts by deepening overall.

Winds allow us to explore other possibilities involving the heat budget. If westerlies blow over the basin instead of easterlies, a very different solution results as shown in Fig. 6. The structure and the circulation are in many ways entirely reversed. The overturning streamfunction now shows convergence on the equator, equatorward flow at the surface, and poleward flow in the subsurface. Many of the features are symmetrically preserved, and the tilt of the thermocline is opposite to that seen in the previous examples. However, the thermocline is now much deeper and sharper, and the undercurrent is absent because there is no equatorward subsurface flow to feed it.

The heat budget perspective provides a rationale for these results, independent of specific dynamical arguments. The same thermal forcing is applied, but cold water is now flowing poleward and upwelled at high latitudes. It is then warmed in its surface equatorward flow and then converged at the equator, where the thermocline is a boundary layer between Ekman convergent flow and diffusively driven upwelling. The warming now occurs at high latitudes, but to balance the global heat budget the model must also cool the same water, and it can only do so at high latitudes where the restoring temperatures are cold. As a result, warm water returns poleward in boundary currents and is cooled at the same latitude at which it is warmed. The heat budget is therefore satisfied locally, no heat transport occurs, and the thermocline is deep as no equatorward outcropping is necessary.

While the strength of the forcing is identical to the easterly case, the solution is very different because of the dramatically different effects of the circulation on the heat budget.

d. Sensitivity to tropical damping

The global heat budget alone does not constrain uniquely the depth of the thermocline. It only does so given a number of externally imposed factors. The atmosphere is here represented as a restoring boundary condition toward a given temperature: the atmosphere has infinite heat capacity, an assumption that is obviously unacceptable and will have to be eventually relaxed by using a coupled model. However, even in this simplified context, the atmospheric role is parameterized

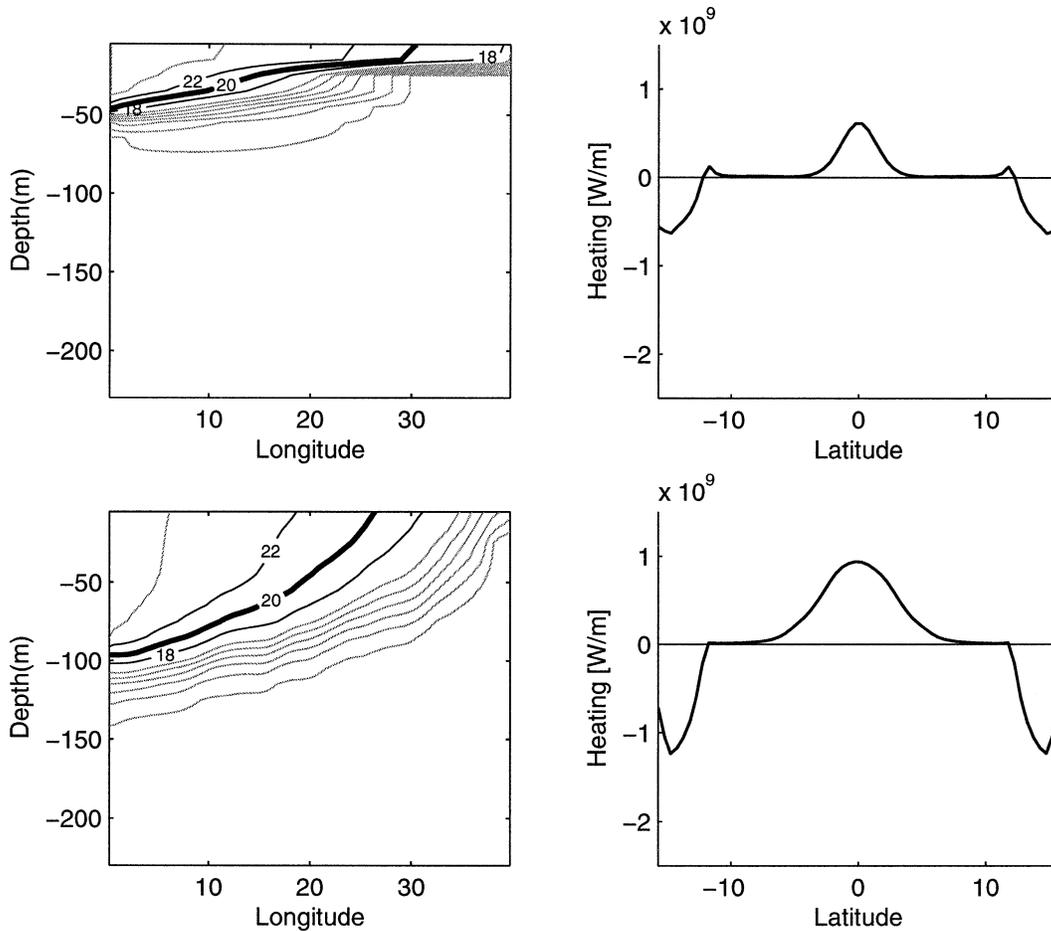


FIG. 5. (right) Zonally integrated heat fluxes, and (left) corresponding thermal structure at the equator for different values of wind strength: (top) $\tau = 0.25 \times 0.05$ Pa, and (bottom) $\tau = 1 \times 0.05$ Pa.

through an adjustable parameter, α , which represents the capacity of the atmosphere to absorb heat for a given temperature difference. In the cases shown up to now this parameter is constant and the same everywhere.

There is no reason in principle for this parameter not to change with location. An example of such a case is given in Fig. 7. Here two cases are compared, the first being the default case with $\alpha = 50 \text{ W m}^{-2} \text{ K}^{-1}$ everywhere [see Eq. (6)]. The other case is one in which α is $50 \text{ W m}^{-2} \text{ K}^{-1}$ in high latitudes and then decreases cosinusoidally to $10 \text{ W m}^{-2} \text{ K}^{-1}$ at the equator.

Because nothing changes in high latitudes—the wind strength and the restoring strength are all the same—the cooling region is unmodified. The upwelling region, and therefore the thermocline structure, are, however, different. Because the heat loss is the same, the same heat gain must occur. However, the strength of the restoring is now up to 5 times as weak in the region of upwelling. This entails that the size of the cold tongue, and the temperature difference between the surface and the restoring, must increase. Such an increase can only be accomplished, for fixed winds, by a shoaling of the

thermocline, and this is precisely what happens in this experiment.

This is an example of how the heat budget alone, through the thermal boundary condition, can profoundly modify the thermal structure of the ocean. We must expect any theory for the thermal structure of the ocean to explain such behavior. In the next section we will attempt to construct a theory that takes into explicit account all these aspects of the heat budget of the ocean.

5. Depth of the thermocline: A simple thermodynamic analysis

The experiments of the previous section show how parameters external to the model ocean control the structure and depth of the thermocline through their effect on the heat budget. In this section we shall try to construct a simple theoretical model to describe these effects.

Figure 8 shows an idealized representation of the thermodynamics at work in this system. The ocean is subdivided into two domains: a tropical box and a high-

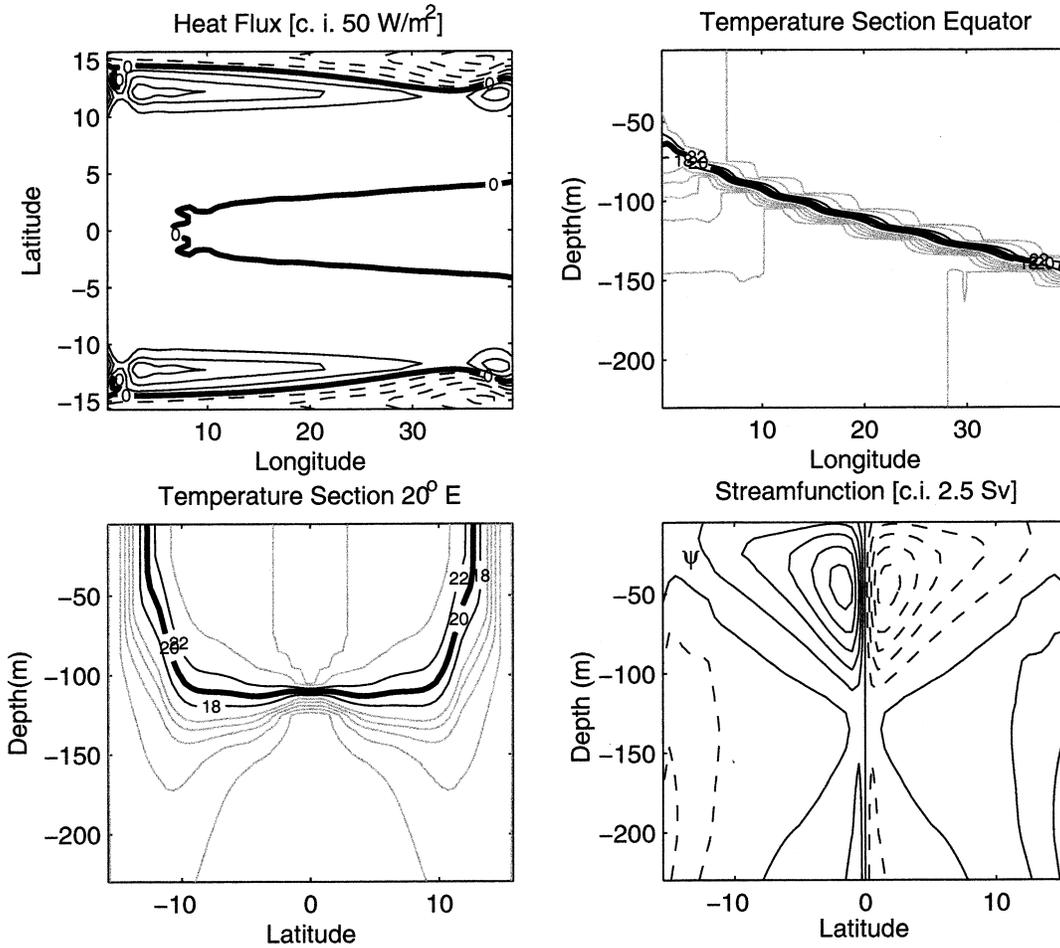


FIG. 6. As in Fig. 3 but for westerlies case.

latitude box. In each box the heat flux is determined by the difference in temperature between the surface of the ocean and the overlying atmosphere; that is,

$$Q^+ = \alpha(T_a - T) \quad \text{for the tropical box and}$$

$$Q^- = \alpha(T_b - T_c) \quad \text{for the high latitude box,} \quad (7)$$

where T is the surface temperature at the Tropics, T_c is the surface temperature in high latitudes, T_a is the specified tropical atmospheric temperature, and T_b is the specified atmospheric high-latitude temperature. We assume that the high-latitude temperature T_c is also the temperature at the base of the thermocline, and therefore of the abyss. By subtracting the two equations in Eq. (7) we get an expression for the static stability of the ocean $\Delta T = T - T_c$ as a function of the heat budget:

$$\Delta T = \Delta T^* - \frac{1}{\alpha}(Q^+ - Q^-), \quad (8)$$

where $\Delta T^* = T_a - T_b$. Equation (8) expresses the important fact that the static stability of the ocean cannot be imposed from the outside as an external parameter, but rather is calculated as part of the solution. In the

limit of $\alpha \rightarrow \infty$, $\Delta T = \Delta T^*$ and the surface temperatures are imposed. In general, though, that is not going to be the case.

The tropical and high-latitude regions are connected by a poleward flow, comprising two components. The first is an Ekman transport v_{Ek} , directly related to the wind strength τ , as $v_{\text{Ek}} = \tau/(f h)$, where h is the depth of the mixed layer within which the Ekman flow occurs. This is the wind-driven flow, and by continuity there will be a vertical flow upwelling at the equator of strength $w_{\text{Ek}} = v_{\text{Ek}} h/L$, where L is the horizontal scale of the upwelling region.

The second part of the flow connecting the subtropics to the Tropics is a response to the meridional density gradient, resulting in a meridional boundary flow v_g . The strength of this flow needs to be parameterized. We shall assume that it is simply a geostrophic current in the meridional direction, the strength of which is given by $v_g = g \gamma D \Delta T / (f L)$: D is the depth of the thermocline and is the depth at which a boundary layer separates the cold abyssal waters from the warm surface tropical waters. This assumption is justified by the work of Park and Bryan (2000).

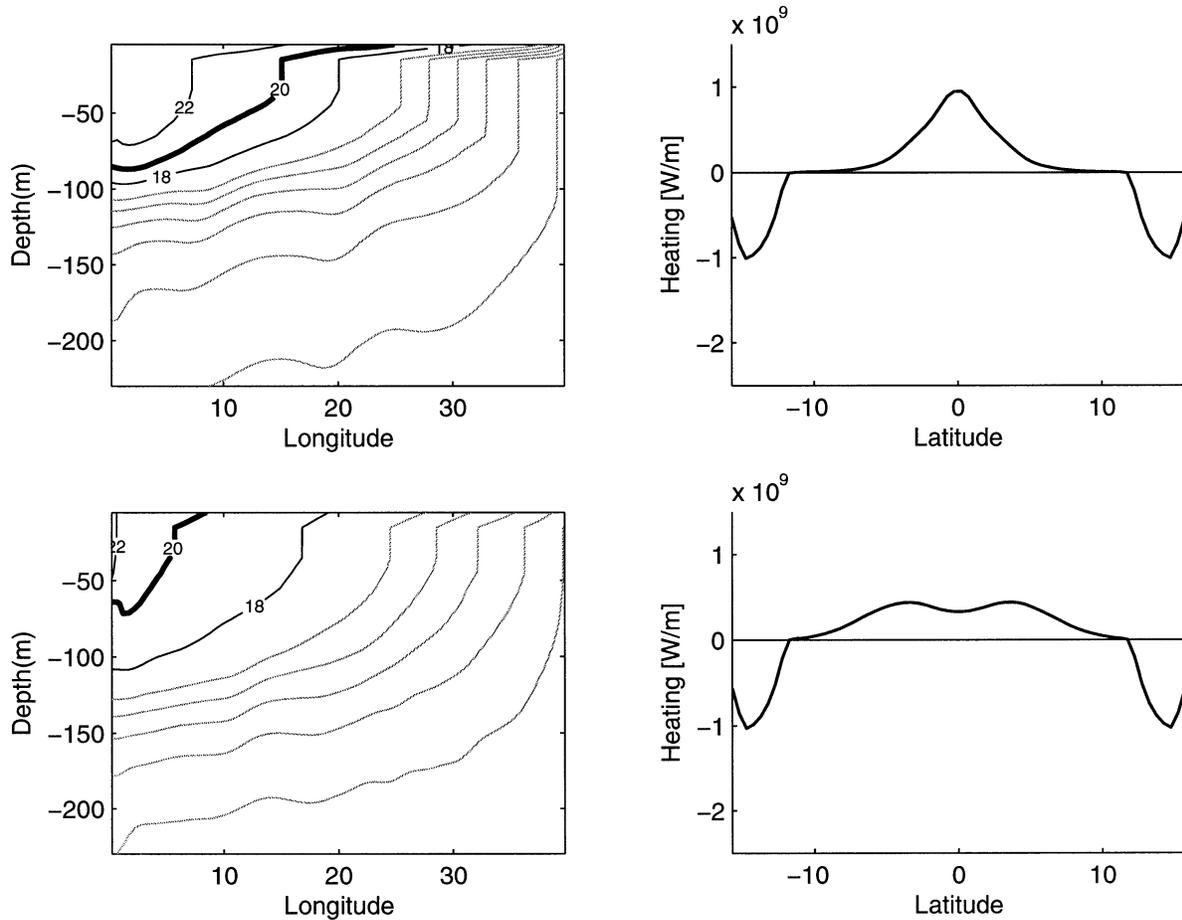


FIG. 7. (right) Zonally integrated heat fluxes, and (left) corresponding thermal structure at the equator. Comparison of mean solution with different tropical restoring coefficients (see text): (top) default case with constant restoring of $50 \text{ W m}^{-2} \text{ K}^{-1}$ and (bottom) case in which the restoring strength decreases to $10 \text{ W m}^{-2} \text{ K}^{-1}$ at the equator.

We now need independent expressions for Q^+ and Q^- . In the Tropics the integrated surface heating Q^+ is partly transferred through diffusion to the abyss and

partly taken up by the conversion of water in the wind-driven flow:

$$Q^+ = L^2 \kappa \frac{\Delta T}{D} + L^2 w_e (T - T_s). \quad (9)$$

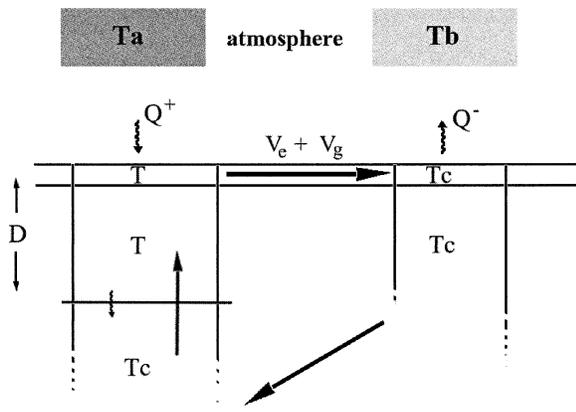


FIG. 8. Graphical representation of the model in section 5. See text for explanation.

The first term on the rhs is a crude estimate of the heat that reaches the abyss through diffusion. The second term on the rhs of Eq. (9) is the heat taken up by the wind-driven overturning. The strength of the volume transport is $w_{ek} L^2$. Notice, however, that the flow will not encounter a temperature difference ΔT but rather on average $T - T_s$, where T_s is the average temperature of the water at the base of the mixed layer of depth h . This reflects the fact that what matters for the heat transport in the tropical box is not the overall static stability of the ocean, but rather the temperature of the water entrained into the mixed layer. During El Niño, for instance, the thermocline flattens in the east Pacific, and the cold water at depth is invisible to the surface entrainment. The static stability has not changed, but the temperature of the water entrained into the mixed layer

has. This fact reflects the dependence of the tropical heat budget on the depth of the thermocline and not just on the temperature difference across the thermocline. We choose as a simple model of such a dependence:

$$T_s = T - \frac{(T - T_c)h}{D} \quad (10)$$

so that, when $D = h$, the water entrained into the mixed layer is at a temperature T_c ; when $D \rightarrow \infty$, the entrained water is at a temperature T . Substituting Eq. (10) into Eq. (9) we get

$$Q^+ = L^2(\kappa + w_e h) \frac{\Delta T}{D}, \quad (11)$$

which relates the tropical heat budget to the depth of the thermocline.

The heat loss in high latitudes Q^- is balanced by advection of warm water by the oceanic transport. We have assumed that $v = v_{\text{Ek}} + v_g$ and that v_{Ek} extends vertically over a depth h , while v_g extends to a depth D . Therefore,

$$Q^- = -v_{\text{Ek}} \Delta T h L - g\gamma \frac{D^2(\Delta T)^2}{f}, \quad (12)$$

where the first term is due to the Ekman transport and the second term is due to the geostrophic transport.

If we now assume that the heat budget is balanced, that is, that

$$Q^+ = -Q^-,$$

then we get from Eqs. (8), (11), and (12) a system of two equations in two unknowns, D and ΔT :

$$v_{\text{Ek}} \Delta T h + g\gamma \frac{D^2(\Delta T)^2}{fL} = \frac{L}{D}(\kappa + w_{\text{Ek}} h) \Delta T \quad \text{and} \\ \Delta T = \Delta T^* - \frac{2\Delta T}{D\alpha}(\kappa + w_{\text{Ek}} h). \quad (13)$$

Equations (13) constitute our theory for the structure of the thermocline. Let us now consider the results of previous theories.

It has been argued that the adiabatic solution can be interpreted as the nondiffusive limit of a diffusive solution (Young and Ierley 1986; Salmon 1990; Samelson and Vallis 1997). Diffusion is then the only process responsible for setting the background stratification, as proposed by Tziperman (1986). In that case the only thermodynamic component of the system should be diffusion and for the purposes of our thermodynamic budget the wind-driven circulation should not cross temperature gradients (i.e., $v_e, w_{\text{Ek}} \Delta T = 0$). In this case Eq. (13) reduces to

$$D = \left[\frac{f^2 L \kappa}{(g\beta\gamma\Delta T)} \right]^{1/3} \quad \text{and} \\ \Delta T = \Delta T^* - \frac{2\Delta T}{D\alpha} \kappa, \quad (14)$$

where $\beta \sim f/L$. This recovers the advective diffusive estimate of Eq. (3). It is a special case in which the diffusively driven overturning circulation is responsible for the heat transport. This is the limit postulated in the work of Tziperman (1986) (see also Walin 1982; de-Szoeke 1995). Notice that we also obtain an estimate for the static stability ΔT , which is expected to differ from ΔT^* by a factor proportional to diffusivity.

In the limit of very small diffusion the assumption of a diffusive thermocline poses serious limitations to the heat transport that can be effected by the ocean. While the open-ocean diffusivity can be very small (Ledwell et al. 1993), mixing in the surface mixed layer is expected to be high so that, as open-ocean diffusion decreases, the wind-driven circulation must dominate the heat transport. If open-ocean diffusion κ is identically zero, then Eqs. (13) reduce to

$$v_{\text{Ek}} \Delta T h + g\gamma \frac{D^2(\Delta T)^2}{fL} = \frac{L}{D} w_{\text{Ek}} h \Delta T \quad \text{and} \\ \Delta T = \Delta T^* - \frac{2\Delta T}{D\alpha} w_{\text{Ek}} h. \quad (15)$$

In the absence of any response to the density gradient ($v_g = 0$) the solution becomes a trivial one and $D = Lw_{\text{Ek}}/v_{\text{Ek}} = h$, the mixed layer depth. This simply means that all water that is converted to cold temperatures must be returned warm by being exposed to the surface. In this limit the temperature difference ΔT is irrelevant to the depth of the thermocline, because for any ΔT the same mass transport will cross the temperature gradient twice, once through being warmed and then through being cooled. When v_g is not zero, the solution is not trivial. The entrainment into the tropical mixed layer must balance the heat loss associated with both midlatitude Ekman flow and boundary current. For a deep thermocline, D will behave as $D = [fL^2 w_{\text{Ek}} h / (g\gamma\Delta T)]^{1/3}$.

In the limit of no diffusion the wind-driven circulation is entirely responsible for determining the background stratification. This limit provides a closure for the ventilated thermocline problem that depends only on the winds and on the properties of the mixed layer. This limit contradicts the idea that the wind-driven circulation is just a finite-amplitude perturbation on a background stratification determined by mixing. Rather, in the limit of small diffusion the wind-driven circulation determines its own background stratification and diffusion is just a perturbation of that.

For the heat transport across a circle of latitude to be zero, the Ekman transport must be opposite in direction to the geostrophic transport so that

$$g\gamma \frac{D^2 \Delta T}{fL} = -v_{\text{Ek}} h. \quad (16)$$

This limit gives a prediction for the depth of the thermocline, $D = [-fLv_{\text{Ek}}h/(g\gamma\Delta T)]^{1/2}$, which is akin to the critical depth for the eastern boundary depth derived by Pedlosky (1987). In our simulations, this balance is

realized in the westerly wind case (Fig. 6), and is possibly true for a model such as that of Samelson and Vallis (1997) in which very little heat transport is effected because the mass transport in the boundary current is almost entirely recirculated equatorward in the horizontal gyre.

6. Parameter dependence of the theory

Figure 9 shows how the depth of the thermocline D (top), the static stability ΔT (middle), and the heat transport H (bottom) depend on the externally imposed parameters τ and ΔT^* for different values of diffusivity. Here H is defined in terms of Q^+ , the heat gained in low latitudes:

$$H = \rho_0 C L^2 (\kappa + w_e h) \frac{\Delta T}{D}, \quad (17)$$

where C is the heat capacity of water and ρ_0 is its density. The two terms on the right-hand side in this equation represent respectively the contribution of the deep thermohaline circulation, whose intensity depends on κ , and the wind-driven circulation.

For low diffusivity ($\kappa = 10^{-6} \text{ m}^2 \text{ s}^{-1}$), the depth of the thermocline is almost insensitive to the imposed temperature gradient ΔT^* and is rather a function of the wind strength. For weak winds the static stability is determined by the imposed temperature gradient, while for strong winds it deviates from ΔT^* . The heat transport also depends mostly on wind strength, although sensitivity to ΔT^* , through its effects on D and ΔT , is also present.

As diffusion is increased, the sensitivity to the wind forcing is reduced and that to the imposed temperature difference increases. This is a result of the increased role of the thermohaline circulation in the thermodynamics of the system as that component of the circulation responds to the temperature gradient. The thermocline depth D is more sensitive to ΔT^* , and the deviation of the static stability from the imposed temperature gradient increases. This change is reflected in the heat transport, which becomes more sensitive to the imposed temperature gradient and less sensitive to the wind strength.

Last, for very high diffusion ($\kappa = 10^{-2} \text{ m}^2 \text{ s}^{-1}$) the solution is almost insensitive to the winds and thermohaline effects dominate. The thermocline is very diffuse and deep, and the static stability of the system is far from the imposed gradient ΔT^* . The heat transport is effectively independent of wind strength as most of it is effected by the diffusive thermohaline circulation.

In classic advective diffusive theory, the heat transport depends on κ through D alone, because ΔT is taken to be a constant. That leads to a $\kappa^{2/3}$ dependence of the heat transport on diffusivity (Park and Bryan 2000). Modeling studies vary, in their simulations of heat transport sensitivity, from $\kappa^{2/3}$ (e.g., Vallis 2000) to $\kappa^{1/2}$ (Marotzke 1997). Recently Park and Bryan (2000) found that

a $\kappa^{2/3}$ dependence is always recovered if the variation of ΔT with κ is also taken into account. Equations (13) can be solved numerically to show that, for high κ , $\Delta T^* - \Delta T \sim \kappa^{1/3}$. This is the empirical dependence found by Park and Bryan (2000) and also confirmed in our simulations.

These results are in good agreement with the changes in thermal structure obtained with our numerical general circulation model and shown in Figs. 4 and 5.

7. Discussion and conclusions

The results presented here demonstrate that the constraint of a balanced heat budget at the ocean surface strongly influences the thermal structure of the ocean. The depth of the tropical thermocline is such that the oceanic heat uptake in the upwelling zones of low latitudes balances the heat loss in mid- and high latitudes. The relationship between the heat budget and depth of the thermocline in the Tropics provides a conceptual closure for the background stratification invoked by theoretical models of the ventilated thermocline. It furthermore allows us to put in a unified context the previous conceptual models of the thermocline, which are shown to belong to particular limits of this theory.

The abyssal and intermediate circulations were not realistically simulated in these runs. The absence of high-latitude convection and deep-water formation limits the extent to which deep processes can alter the upper ocean. In the presence of deep-water formation, heating must be invoked to close the deep circulation. Where and how this heating occurs are still a matter of debate, but it is unlikely that such questions can be answered within the context of basin simulations alone.

In this study, the oceanic circulation that connects the regions of heat loss and heat gain has two main components: the deep thermohaline and the shallow, wind-driven circulation of the ventilated thermocline. Their relative importance depends on the oceanic diffusivity. Low diffusivities, which favor the wind-driven circulation, maximize the importance of upwelling zones in the oceanic adjustment to a change in the heat budget. High diffusivities minimize the role of upwelling zones because oceanic heat uptake occurs over a region with a large areal extent. Observations suggest that, at least in the main thermocline, mixing is indeed small (Ledwell et al. 1993), including that the observed world is one in which upwelling zones are of paramount importance. The choice of simulating a thermohaline circulation by specifying a mixing rate is driven by the necessity of simplifying the problem. Clearly this is a limited representation of thermohaline effects in this study (Wunsch 2002). However, in the absence of a satisfying closure for the observed mixing in the ocean, it does allow a useful comparison with the behavior of most existing models.

The results presented here depend critically on an oceanic surface boundary condition of the form given

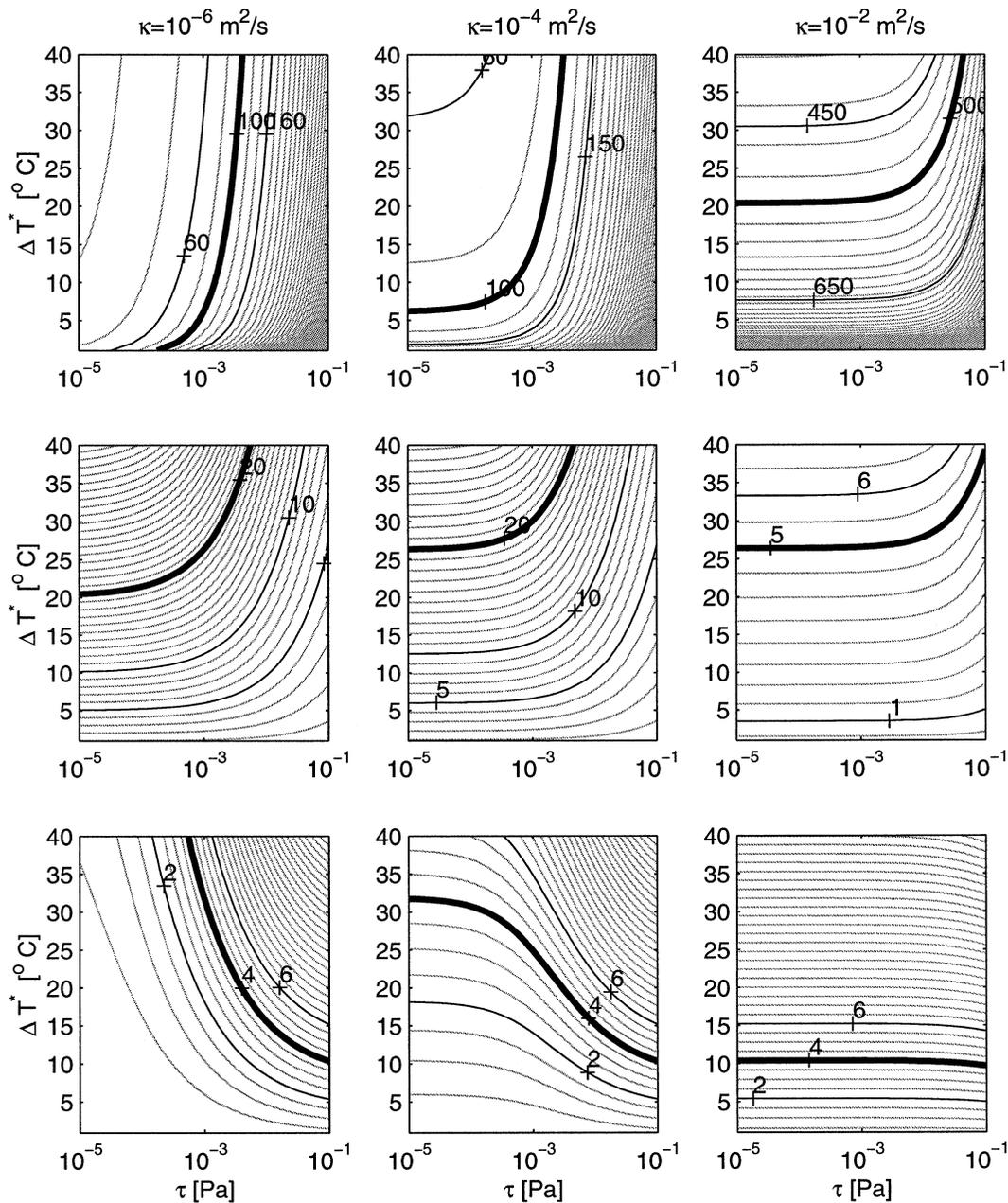


FIG. 9. (top) Depth of the thermocline D (m) for three different values of diffusivity (κ increases left to right). (middle) Static stability ΔT^* ($^{\circ}\text{C}$) for the same three values of diffusivity. (bottom) Heat transport H (10^{14} W). Parameters are shown as functions of imposed wind strength τ and imposed temperature atmospheric difference ΔT^* .

by Eq. (6). In that equation the specified temperature T^* represents atmospheric temperatures, and T is the oceanic surface temperature, which is determined by oceanic processes. The surface winds are specified independently. In reality and in coupled ocean–atmosphere models, the winds and T^* are related and depend on T . The purpose of the approach taken here is to explore the effect of different relations between T^* and the winds, thus shedding light on what could happen in worlds different from the present, familiar one and on

what happens in different coupled ocean–atmosphere models. For example, in some of those models the ocean is simply a mixed layer so that $T = T^*$ at each point, and the constraint of a balanced heat budget is met locally at each point. In such models horizontal heat transport is zero, and there are no upwelling zones that can adjust in response to nonlocal changes in the heat budget. In the case of coupled ocean–atmosphere models in which the oceanic component has low horizontal resolution, diffusivity is necessarily large and the roles

of the wind-driven circulation, and of the upwelling zones in particular, are minimized as mentioned above. Last, some models have, in Eq. (17), an additional, specified term Q^* that is usually referred to as a flux correction term. The presence of such a term inhibits the ability of the oceanic thermal structure, and heat transport, to adjust to a change in the oceanic heat budget.

With the previously illustrated limitations in mind, it is possible to speculate on the implications of this theory for climate change, as it does make predictions for the heat transport and heat content of the ocean under different surface conditions. For example, an increase in the oceanic heat loss, as is likely in glacial climates, results in a shoaling of the thermocline so that increased heat uptake can balance the loss. At the same time poleward heat transport increases. In a warm world with little oceanic heat loss, on the other hand, the tropical thermocline is likely to be deep so that the oceanic heat gain can be correspondingly small; poleward heat transport is reduced.

Today the thermohaline circulation and wind-driven circulation make approximately comparable contributions to the poleward heat transport. Does that partitioning remain the same in a different climate? In his study of the partitioning of poleward heat transport between the ocean and atmosphere, Held (2001) assumes that the oceanic component can be expressed as $Q = V\Delta T$. The volume transport V was assumed to be set by the winds, placing that work in the limit of no diffusion. The surface temperatures and vertical stability ΔT were also assumed to be fixed, leading to a heavily constrained partitioning. In this work we have shown how the surface temperature gradient is a dynamic quantity that depends on the circulation itself. The limit proposed by Held is therefore one in which $\alpha \rightarrow \infty$. Furthermore, because the heat budget depends not only on ΔT but also on the depth of the thermocline D , the constraint proposed by Held might be relaxed. Consideration of the fully coupled problem is necessary to address the issue.

Geometrical effects such as the presence of an open channel or the effects of a larger basin were ignored, as we limited ourselves to a small closed basin. Additional effects on the stratification below the tropical thermocline are expected from the inclusion of the Southern Ocean (e.g., Vallis 2000; Klinger et al. 2003) and from the inclusion of the effects of eddies (e.g., Marshall et al. 2002). Gnanadesikan (1999) proposed a scaling for the stratification of the ocean intended to take into account these effects. That scaling, based on a mass budget rather than on the surface heat budget, does not take into account the role of tropical upwelling, which we have shown to be critical for closing the heat budget of the ocean and for determining the depth of the tropical thermocline. It is, however, appropriate for the stratification below the tropical thermocline and for describing the sensitivity of the overturning circulation in the Atlantic (Klinger et al. 2003). The two theories are

therefore complementary, and attempts to combine these two conceptual frameworks in a unified picture are currently under way.

Acknowledgments. The authors thank Robert Hallberg, Geoff Vallis, Kirk Bryan, Isaac Held, and Anand Gnanadesikan for useful comments at different stages of this work. Author GB gratefully acknowledges NASA Fellowship NGT5-30333, NOAA Grant NA16GP2246, NASA Grant NAG5-12387, and NASA, from JPL Contract 1229837, are gratefully acknowledged.

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