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[1] One of the most important contributions the ocean makes to Earth's climate is through its poleward heat transport: about 1.5 PW or more than 30% of that accomplished by the ocean-atmosphere system (Trenberth and Caron, 2001). Recently, concern has arisen over whether global warming could affect this heat transport (Watson et al., 2001), for example, reducing high latitude convection and triggering a collapse of the deep overturning circulation (Rahmstorf, 1995). While the consequences of abrupt changes in oceanic circulation should be of concern, we argue that the attention devoted to deep circulations is disproportionate to their role in heat transport. For this purpose, we introduce a heat function which identifies the contribution to the heat transport by different components of the oceanic circulation. A new view of the ocean emerges in which a shallow surface intensified circulation dominates the poleward heat transport. **Citation:** Boccaletti, G., R. Ferrari, A. Adcroft, D. Ferreira, and J. Marshall (2005), The vertical structure of ocean heat transport, *Geophys. Res. Lett.*, *32*, L10603, doi:10.1029/2005GL022474.

1. Introduction

[2] Model simulations suggest that in the absence of oceanic heat transport the high latitudes would cool significantly, polar caps would spread equatorwards, and the Earth would freeze over [*Winton*, 2003]. While global ocean heat transport can be deduced from estimates of air-sea fluxes [*Trenberth and Caron*, 2001], its partitioning among different components of the circulation is difficult to determine from observations or models, and remains a source of debate.

[3] Traditional views of the ocean assign the bulk of the heat transport to the deep overturning circulation [*Roemmich and Wunsch*, 1985]. More recently *Talley* [2003] has attempted to reconstruct the vertical structure of the oceanic circulation from data by identifying water mass ventilation for shallow, intermediate, and deep waters. She concludes that shallow overturnings dominate the transport in the Pacific, but not in the Atlantic, where the deep overturning carries the bulk of the heat. Despite current efforts to accurately constrain the vertical structure of the oceanic heat transport, a view of the ocean circulation in which the deep circulation is responsible for most of the heat transport pervades the literature, especially when concerning climate change [*Rahmstorf*, 1995; *Alley et al.*, 2003]. A clear way of diagnosing the vertical structure of the oceanic heat transport is still lacking.

[4] For a circulation flowing poleward near the surface, where temperatures are high, and returning equatorward at depth, where temperatures are low, the poleward heat transport is proportional to the strength of the circulation multiplied by the temperature difference encountered. The mean absolute temperature of the circulation is irrelevant as it just represents heat being carried around without being exchanged with the atmosphere. By calculating such a quantity for each circulation, it is possible to compare the contribution each one makes to the poleward heat transport. This paper develops the tools to make such calculations.

2. The Model

[5] The model used in this study is the MIT ocean general circulation model with a horizontal resolution of 2.8°, and 15 vertical levels [*Marshall et al.*, 1997]. At this coarse resolution processes such as convection, mixing, and transfer of properties by the mesoscale eddy field, are parameterized. The surface heat and fresh water fluxes are each the sum of two terms: an imposed flux and a restoring term proportional to the difference between surface values and climatology [*Jiang et al.*, 1999]. The restoring timescales are 60 and 90 days for temperature and salinity respectively. The model is therefore capable of modifying its surface fluxes in response to changes in circulation. The model is initialized from Levitus climatology, forced by monthly mean climatological surface wind stresses [*Trenberth et al.*, 1990], and run for 7000 years until it has reached equilibrium. A constant diapycnal mixing coefficient κ of 0.3×10^{-4} m^2/s is used.

[6] Figure 1a shows the stream function ψ (where $\partial_y\psi = \overline{w}$, $\partial_z\psi = -\overline{v}$, and $\overline{v}$ and $\overline{w}$ are the zonally integrated meridional and vertical velocities respectively) for the global ocean model used here. Shallow cells, confined to the first 500 m, are connected to the processes of midlatitude subduction and equatorial upwelling. Deep cells, occupying the much vaster abyss, are connected to high latitude convection and abyssal mixing. It is the potential collapse of these cells – as a result of the shut down of North Atlantic Deep Water formation for example – that is believed to trigger major climate shifts, because of the sudden decrease in meridional heat transport by the ocean [*Alley et al.*, 2003].

3. The Heat Function

[7] A common way to diagnose the heat transport is to compute the vertical integral of the temperature flux vT across an east-west section bounded by continents [*Bryden and Imawaki*, 2001]. This ensures that, as long as the total mass flux across the section is zero, the heat transport is due

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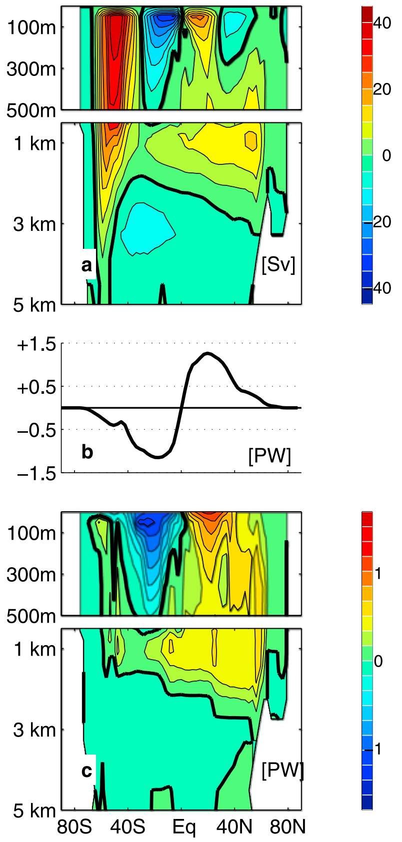

Figure 1. (a) Stream function. Meridional volume transport in Sv (10^6 m^3/s) inferred from the global circulation model (ψ is equivalent to mass transport as seawater is nearly incompressible). Red shading is clockwise and blue is counter-clockwise flow. Contour interval is 5 Sv. (b) Total heat transport of the global circulation model. (c) Heat function. Meridional heat function in PW (10^{15} W) inferred from a global circulation model. Red shading is northward heat transport and blue is southward heat transport. Contour interval is 0.17 PW.

to water flowing northward at one temperature and returning at a different temperature. However this approach does not distinguish between the contribution of different circulations that make up the meridional overturning flow.

[8] To address this issue we introduce a new quantity called the heat function, representing oceanic pathways of heat contributing to the total oceanic meridional heat transport. The zonally integrated heat transport in the ocean interior comprises advective and diffusive components:

$$\mathcal{H} = \overline{\mathbf{v}T} - \kappa\nabla\overline{T}, \tag{1}$$

where κ is the temperature diffusivity (details of the following calculations are given in Appendix A.) In the steady state the divergence of $\mathcal{H}$ is zero because there are no sources or sinks of heat in the ocean interior; the only sources are at the surface and are responsible for setting the value of $\mathcal{H}$ there. It is therefore possible to represent the heat fluxes in terms of a scalar function ϕ by defining $\partial_z\phi = \overline{vT} - \kappa_H\partial_y\overline{T}$ and $\partial_y\phi = -\overline{wT} + \kappa_V\partial_z\overline{T}$($\kappa_H$ and κ_V are the horizontal and vertical diffusivities respectively). We have chosen the sign to ensure that the heat function is positive for northward heat transport. However what contributes to the net meridional heat transport is only the component of the heat flux that varies along closed streamline loops: in the zonal mean, heat that is carried poleward at one depth and returned equatorward at another depth does not contribute any net meridional heat exchange. Therefore, we must eliminate from ϕ the portion of advective fluxes along each closed streamline tube that carries fluid at constant temperature. That portion is given by the product of the circulation **v** times a temperature uniform along the flow $\hat{T} = \hat{T}(\psi)$, where ψ are the appropriate streamlines. There is arbitrariness in the choice of $\hat{T}$, and several were tried, but as long as $\hat{T}$ represents a reasonable average temperature for the fluid moving along the streamlines, the results do not qualitatively change. We chose the temperature field $\hat{T}$ which minimizes the total advective heat transport in a least square sense, that is we minimize $\mathbf{v}(T - \hat{T})$ when averaged between two streamlines, as defined in equation (A3). We can therefore define the heat function ϕ as,

$$-\mathbf{i} \times \nabla\phi = \mathcal{H} - \mathbf{v}\hat{T}(\psi). \tag{2}$$

where **i** is a unit vector in the zonal direction.

4. Global Heat Transport

[9] With this definition of heat function it is possible to identify, in the meridional plane, the contribution made by different components of the circulation to the total heat transport. The surface value of ϕ equals the vertically integrated total meridional heat transport. Regions where the heat function is zero do not contribute to the vertically integrated meridional heat transport. Figure 1b shows the vertically integrated meridional heat transport and Figure 1c shows the heat function ϕ for the model run described above. The heat function captures a stronger asymmetry between surface and abyssal flows: it peaks at 1.5 PW at the surface and there is no contribution to the heat transport below 2000 m. Virtually all of the heat transport in the southern hemisphere is confined to a shallow circulation in the upper 500 m. In the northern hemisphere a contribution by a deeper circulation is still present and represents the contribution of NADW. However it only amounts to 0.4 PW out of the total, the rest being surface intensified. The heat function is antisymmetric about the equator and confined between 50°N and 50°S, therefore matching the total heat transport. Heat travels along continuous pathways that connect the midlatitude subduction sites to regions of equatorial upwelling.

[10] Recent analyses of observational data suggest that the contribution of the shallow circulation to the global heat transport is indeed much larger than previously thought, and overall dominates the heat transport [*Talley*, 2003].

5. Atlantic Heat Transport

[11] The deep meridional overturning circulation in the Atlantic, primarily associated with North Atlantic Deep Water formation, often is argued [*Bryan*, 1962] to dominate the oceanic heat transport because it flows poleward at the surface and equatorward at depth, therefore encountering a

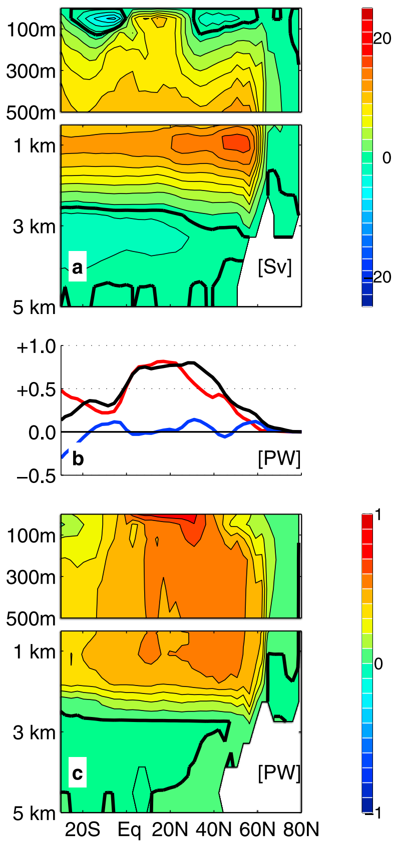

Figure 2. (a) Mass stream function for the North Atlantic. Contour interval is 2 Sv. (b) Decomposition of the advective heat transport $\int \overline{vT}dz$ (black line) into a vertical contribution $\int \overline{v}\overline{T}dz$ (red line) and a horizontal contribution $\int \overline{v'T'}dz$ (blue line). Here $\overline{()} = \int ()dx$ and $()' = () - \overline{()}$. (c) Heat function for the North Atlantic. Contour interval is 0.1 PW.

large top to bottom vertical temperature difference. This is in opposition to the horizontal transport in gyres driven by the winds which, while of the same order of magnitude as the meridional overturning, encounter only the much smaller east-west temperature difference. Therefore, the argument goes, most of the heat transport must be due to the meridional overturning associated with deep water formation. The stream function for the Atlantic is shown in Figure 2a and the traditional decomposition into horizontal and vertical heat transport [*Bryden and Imawaki*, 2001] is shown in Figure 2b. At 24°N this decomposition assigns less than 0.1 PW to the horizontal circulation and 0.8 PW to the vertical circulation. Figure 2a further suggests that the main contribution is due to the deep overturning. However Figure 2c shows that such heat transport is again surface intensified. Both the deep circulation and the wind-driven circulation contribute to the vertical component, the former contributing about 0.4 PW, the latter about 0.5 PW. This is roughly consistent with the analysis by *Talley* [2003], who also recognizes the large contribution of the shallow circulation.

6. Abyssal Mixing and Heat Transport

[12] The importance of the deep overturning circulation in transporting heat is invoked to motivate the study of abyssal mixing in the ocean: it is often argued that the oceanic poleward heat transport is limited by the ability of the ocean to mix dense bottom waters across the stratification of the abyss [*Munk and Wunsch*, 1998].

[13] Figure 1c suggests that the problem of determining the oceanic heat transport is largely independent from that of the abyssal stratification, and that studies of deep abyssal mixing should be seen in this context. The mass transport is sensitive to mixing, but the heat transport also depends on the temperature difference, which is small in the abyss. Figure 3a shows the difference in volume transport between a simulation with a constant diapycnal mixing coefficient of $\kappa = 0.3 \times 10^{-4}$ m^2/s and a simulation in which abyssal mixing has been enhanced by increasing it to 1.7×10^{-4} m^2/s below 2,000 m [*Bryan and Lewis*, 1979].

[14] The deep circulation more than doubles in response to the enhanced abyssal mixing, going from about 9 Sv to 21 Sv (Figure 3a). However the associated heat function (Figure 3b), and the total heat transport, remain virtually unchanged even though in principle the model is free to adjust its surface heat flux in response to changes in the circulation. Therefore while the mean mass transport and the deep stratification are indeed controlled by abyssal mixing, the modelled global heat transport is not [see also *Scott and Marotzke*, 2002]. This is not the case when mixing is also allowed to change in the thermocline so as to affect the surface circulation. Studies where mixing is changed across the whole water column, including the upper 500 m, do indeed find that the total heat transport is affected [*Simmons et al.*, 2004].

7. Conclusions

[15] Our analysis has important implications for assessing the role of the ocean in paleoclimate and climate change. We have shown that the surface circulation dominates the heat transport in a model of the global ocean. It is therefore more likely that ocean heat fluxes will respond to changes in atmospheric winds, which drive the shallow overturning circulations associated with midlatitude subduction and equatorial upwelling [*Boccaletti et al.*, 2004], rather than to changes in abyssal mixing and convection that drive the deep circulations. This does not exclude the possibility that the abyssal flow may affect the heat transport indirectly

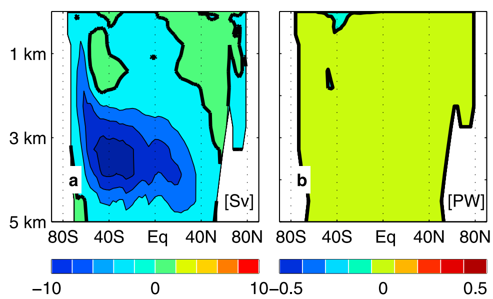

Figure 3. (a) Stream function Anomaly. Difference in meridional volume transport between a run with constant mixing $\times 10^{-4}$ m^2/s and one in which the mixing has been increased to 1.7×10^{-4} m^2/s in the abyss below 2,000 m. Contour interval is 3 Sv. (b) Heat function anomaly. Same as a but for the heat transport. Contour interval is 0.2 PW.

through its effect on the surface flow. However a decrease in deep overturning circulation alone does not imply a significant reduction of the global oceanic heat transport. These results furthermore suggest that variability in the global heat budget transport may be decadal, the timescale of upper ocean variability, rather than centennial or longer, which is the timescale typical of the deep meridional overturning.

Appendix A: Calculation of the Heat Function

[16] The appropriate representation of mass flow for use in computation of the heat function is the residual circulation, which represents both the effects of mean flow and eddies. To proceed we divide the zonal and time mean heat transport in a mean contribution $\overline{\mathbf{v}T}$, a contribution due to the parameterized mesoscale eddy component $\overline{\mathbf{v}_{GM}T} - \overline{K_{Redi}\nabla T}$ (split into advective and diffusive terms as suggested by *Gent and McWilliams* [1990]), and small-scale turbulent fluxes $\kappa\overline{\nabla T}$. The heatflux $\mathcal{H}$ can then be written as,

$$\mathcal{H} = \overline{(\mathbf{v} + \mathbf{v}_{GM})T} - \overline{(K_{Redi} + \kappa)\nabla T}. \tag{A1}$$

In the steady state the heatflux is divergenceless in the ocean interior, $\nabla \cdot \mathcal{H} = 0$, and equal to the surface heat fluxes at the ocean surface.

[17] To obtain the heat function we subtract from the total heatflux the product of the residual circulation, $\mathbf{v}_{res} = \overline{\mathbf{v}} + \overline{\mathbf{v}_{GM}}$, times a temperature which is uniform along residual streamlines $\hat{T} = \hat{T}(\psi_{res})$ (where the residual stream function is given by $\mathbf{v}_{res} = \mathbf{i} \times \nabla\psi_{res}$). Notice that in making this choice we neglect zonal and temporal correlations between $\mathbf{v} + \mathbf{v}_{GM}$ and T fluctuations, which were found to contribute little to the final result.

[18] The temperature field $\hat{T}$ is found by minimizing:

$$\delta \int \left|\left(T - \hat{T}\right)\mathbf{v}_{res}\right|^2 \frac{\mathrm{d}l}{|\nabla\psi_{res}|} = 0, \tag{A2}$$

where dl is the infinitesimal increment along the streamline and δ represents a perturbation in $\hat{T}$. The solution is

$$\hat{T}(\psi_{res}) = \frac{\int |\nabla\psi_{res}|T \; \mathrm{d}l}{\int |\nabla\psi_{res} \; |\mathrm{d}l}. \tag{A3}$$

The flux $\mathbf{v}_{res}\hat{T}$ is by definition constant along streamlines and is the largest portion of the heatflux, in a least square sense, which can be eliminated from $\overline{\mathbf{v}_{res}T}$. It can be shown that subtracting this quantity does not change the vertically integrated heat transport.

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References

Alley, R., et al. (2003), Abrupt climate change, *Science*, *29*, 2005–2010.

Boccaletti, G., R. C. Pacanowski, S. G. H. Philander, and A. Fedorov (2004), The thermal structure of the upper ocean, *J. Phys. Oceanogr.*, *34*, 888–902.

Bryan, K. (1962), Measurements of meridional heat transport by ocean currents, *J. Geophys. Res.*, *67*, 3403–3414.

Bryan, K., and L. J. Lewis (1979), A water mass model of the world ocean, *J. Geophys. Res.*, *84*, 2503–2517.

Bryden, H. L., and S. Imawaki (2001), Ocean heat transport, in *Ocean Circulation and climater*, edited by G. Siedler, J. Church, and J. Gould, pp. 495–474, Elsevier, New York.

Gent, P. R., and J. C. McWilliams (1990), Isopycnal mixing in ocean circulation models, *J. Phys. Oceanogr.*, *20*, 150–155.

Jiang, S., P. H. Stone, and P. Malanotte-Rizzoli (1999), An assessment of the Geophysical Fluid Dynamics Laboratory ocean model with coarse resolution: Annual-mean climatology, *J. Geophys. Res.*, *104*, 25,623–25,645.

Marshall, J., C. Hill, L. Perelman, and A. Adcroft (1997), Hydrostatic, quasi-hydrostatic, and nonhydrostatic ocean modeling, *J. Geophys. Res.*, *102*, 5753–5766.

Munk, W. H., and C. Wunsch (1998), Abyssal recipes II. Energetics of the tides and wind, *Deep Sea Res., Part A*, *45*, 1977–2010.

Rahmstorf, S. (1995), Bifurcations of the Atlantic thermohaline circulation in response to changes in the hydrological cycle, *Nature*, *388*, 825–826.

Roemmich, D., and C. Wunsch (1985), Two transatlantic sections: Meridional circulation and heat flux in the subtropical North Atlantic Ocean, *Deep Sea Res., Part A*, *32*, 619–664.

Scott, J. R., and J. Marotzke (2002), The location of diapycnal mixing and the meridional overturning circulation, *J. Phys. Oceanogr.*, *32*, 3578–3593.

Simmons, H. L., S. R. Jayne, L. C. St. Laurent, and A. J. Weaver (2004), Tidally driven mixing in a numerical model of the ocean general circulation, *Ocean Model.*, *6*, 245–263.

Talley, L. (2003), Shallow, intermediate and deep overturning components of the global heat budget, *J. Phys. Oceanogr.*, *33*, 530–560.

Trenberth, K. E., and J. M. Caron (2001), Estimates of meridional atmosphere and ocean heat transports, *J. Clim.*, *14*, 3433–3443.

Trenberth, K. E., W. G. Large, and J. G. Olson (1990), The mean annual cycle in global ocean wind stress, *J. Phys. Oceanogr.*, *20*, 1742–1760.

Watson, R. T., et al. (Eds.) (2001), *Climate Change 2001: Synthesis Report*, 398 pp., Cambridge Univ. Press, New York.

Winton, M. (2003), On the climatic impact of ocean circulation, *J. Clim*, *16*, 2875–2889.

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