

## The possible role in the ocean heat budget of eddy-induced mixing due to air-sea interaction

Richard J. Greatbatch,<sup>1</sup> Xiaoming Zhai,<sup>1</sup> Carsten Eden,<sup>2</sup> and Dirk Olbers<sup>3</sup>

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[1] The traditional point of view is that in the ocean, the meridional transport of heat is achieved by the wind-driven and meridional overturning circulations. Here we point out the fundamental role played by ocean mixing processes. We argue that mixing (i.e., water mass conversion) associated with eddies, especially in the surface mixed layer, can play an important role in closing the ocean heat budget. Our results argue that the lateral mixing applied at the surface of ocean/climate models should be playing an important role in the heat balance of these models, indicating the need for physically-based parameterizations to represent this mixing.

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### 1. Introduction

[2] The earth exhibits a net radiative gain of heat from the sun in the tropics and middle latitudes, but a net loss in higher latitudes. The earth's fluid envelope is therefore required to redistribute heat from low to high latitudes in order to maintain balance [Gill, 1982]. The ocean is believed to play an important role in this process and the traditional picture is that ocean heat transport is achieved by the circulation [e.g., Bryan, 1991]. In the North Atlantic, it is thought the meridional overturning circulation dominates [Hall and Bryden, 1982; Roemmich and Wunsch, 1985], with warm water flowing northward in the upper part of the water column and colder water returning southward at depth, giving a net northward transport of heat when integrating over the depth of the water column ([Bryan, 1962]). Recently [Boccaletti et al., 2005] have pointed out that the shallow wind-driven overturning cells also contribute to this process. Here we present an alternative point of view and emphasise the fundamental role played by mixing processes in the ocean heat budget. The necessary mixing can arise either from interaction with the atmosphere in the surface mixed layer or from interior mixing. Estimates of the diapycnal diffusivity in the interior of the ocean [Ledwell et al., 1993] are of the order  $10^{-5} \text{ m}^2 \text{ s}^{-1}$ , and we argue that this level of mixing is insufficient to close the heat budget (see sections 2 and 3). Although more

enhanced mixing has been found over rough topography, e.g. [Ledwell et al., 2000], this mixing takes place at too great a depth to affect the heat budget for the top 1 km or so of the ocean where the thermocline is located (see Figure 1). Mixing inferred from water mass transformation theory [e.g., Speer, 1997] also supports the small values found in the ocean interior. Eddies, on the other hand, are known to play an important role in shaping the large-scale ocean circulation, especially in the Southern Ocean [Danabasoglu et al., 1994; Rintoul et al., 2001], but it has not been clear what role the eddies play in the overall heat balance. Early work suggested that eddies do not have a fundamental role to play [Drijfhout, 1994; Bryan, 1996]. In this letter, we combine a new theory [Eden et al., 2007] with observed data to argue, by contrast, that the redistribution of heat by eddies plays a fundamental role. Our approach is an extension of that taken by [Radko and Marshall, 2004] [see also Marshall et al., 2002; Hughes, 2002], except that rather than analyse an idealised model set-up (as in Radko and Marshall [2004]), we apply the theory to ocean data. Importantly, we emphasise the role played by the surface mixed layer and, in particular, we stress a new aspect, namely the importance of mixing that arises from the damping of the sea surface temperature (SST) variance by the surface heat flux [see Zhai and Greatbatch, 2006b]). Our ultimate aim is to show the importance of heat transfer associated with mesoscale eddies in the 3-D heat budget of the ocean. However, to do this we must begin with the heat budget for the ocean after zonal averaging (section 2), for which standing eddies are important, and then make use of the zonally-averaged case to extend the theory to three-dimensions in section 3, where the importance of mesoscale eddies is revealed.

### 2. Zonal Averaging

[3] The potential temperature in the ocean reveals a bowl-shaped warm water pool, with isentropes outcropping at the surface on both sides of the equator (see Figure 1). The question arises as to how the heat input at the surface at low latitudes escapes from the “bowl” and redistributes itself poleward in order to maintain balance. Following zonal averaging at fixed height and time averaging to ensure a statistically steady state, the equation for the ocean heat budget, following [Eden et al., 2007], can be written as:

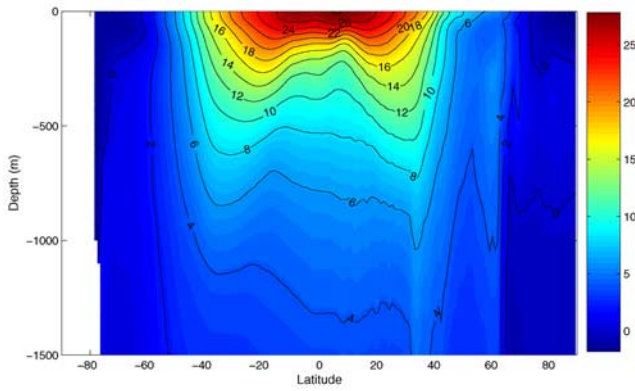
$$\nabla \cdot (L_w \mathbf{u}^* \bar{T}) = \bar{Q} + \nabla \cdot (L_w K_e \nabla \bar{T}) \quad (1)$$

where the overbar denotes the averaging operator (zonal and time averaging),  $T$  is the potential temperature,  $\mathbf{u}^*$  is the “residual mean” velocity (sum of the Eulerian mean and eddy-induced transport velocity; [see Gent et al., 1995],  $L_w$

<sup>1</sup>Department of Oceanography, Dalhousie University, Halifax, Nova Scotia, Canada.

<sup>2</sup>Leibniz Institute of Marine Sciences, University of Kiel, Kiel, Germany.

<sup>3</sup>Alfred Wegener Institute for Polar and Marine Research, Bremerhaven, Germany.



**Figure 1.** Zonally-averaged potential temperature ( $^{\circ}\text{C}$ ) from the World Ocean Atlas 1998 [Levitus *et al.*, 1998].

is the zonal width of the ocean as a function of latitude and depth, and  $Q$  is the total, instantaneous thermal forcing associated, in the surface mixed layer, with the surface heat flux and 3-D turbulence (e.g., leading to entrainment/detrainment etc.) and, in the ocean interior, with microstructure mixing [Gregg *et al.*, 2003].  $K_e$  is the thermal diffusivity associated with the departures from the average. Eden *et al.* [2007] give the following expression for  $K_e$  in statistical steady conditions:

$$K_e |\nabla \bar{T}|^2 = -\overline{Q'T'} + \text{higher order terms} \quad (2)$$

where the prime denotes the departure from the zonal/time average. Noting that, in statistically steady state, the equation for the eddy variance,  $\bar{\phi} = \frac{\overline{T'^2}}{2}$  is

$$\nabla \cdot (L_w \mathbf{u} \bar{\phi}) = L_w [-\overline{\mathbf{u}'T'} \cdot \nabla \bar{T} + \overline{Q'T'}] \quad (3)$$

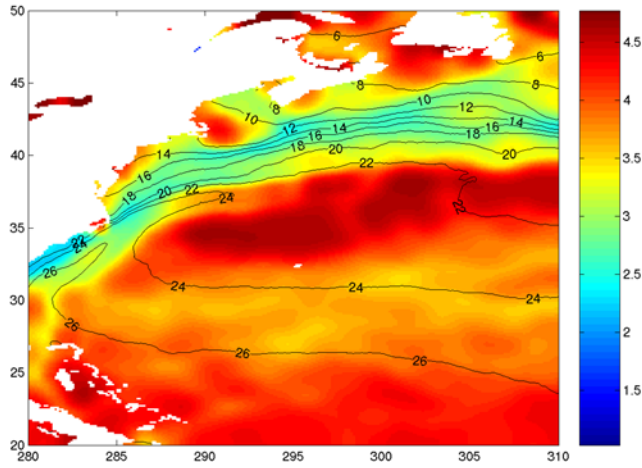
we see that the second term on the right hand side of equation (2) is the same as the  $\overline{Q'T'}$  term in equation (3). The higher order terms in equation (2) arise from rotational fluxes used to absorb the advective flux of variance on the left hand side of equation (3) (see Eden *et al.* [2007] for details) leaving a balance between production,  $-\overline{\mathbf{u}'T'} \cdot \nabla \bar{T}$  and dissipation  $\overline{Q'T'}$  to determine  $K_e$ . Furthermore, the  $\overline{Q'T'}$  term is negative when variance is being dissipated, in which case  $K_e$  emerges as a positive coefficient. (Equations (7) and (11) provide a simple illustration of  $K_e$  when  $Q$  is a Newtonian relaxation term.) We now integrate equation (1) over the area (called the “control volume”) between the sea surface and a mean isentrope, say  $\bar{T} = T_o$ . The residual mean advection term integrates to zero and plays no role in the subsequent balance (using the fact that the normal component of  $\mathbf{u}^*$  is zero at the sea surface and  $\nabla \cdot (L_w \mathbf{u}^*) = 0$ ). Physically, this is because the potential temperature of the water that is advected into the control volume (the “bowl” of warm water) is the same as the potential temperature of the water that is advected out. The resulting balance (see [Walín, 1982]) is therefore

$$\int_s L_w (K_s + K_e) \frac{\partial \bar{T}}{\partial n} ds = \frac{\mathcal{H}}{\rho c_p} \quad (4)$$

where the integral is taken along the isentrope  $\bar{T} = T_o$ ,  $n$  is the coordinate perpendicular to the isentrope, and  $\rho$ ,  $c_p$  are the density of sea water and the specific heat at constant pressure, respectively. In equation (4) we have split the thermal forcing  $Q$  into two parts:  $K_s$  is the diffusivity associated with the 3-D mixing, in particular microstructure mixing in the ocean interior, and  $\mathcal{H}$  is the net input of heat at the surface. Equation (4) says that the heat input at the surface,  $\mathcal{H}$ , is balanced by mixing processes, either the 3-D mixing associated with  $K_s$  acting on the mean gradient, or mixing (i.e., water mass conversion) associated with the eddies, i.e., the departures from the averaging operator, and represented by  $K_e$ . A striking feature of equation (4) is that there is no appearance of the mean circulation associated with the wind-driven or meridional overturning circulations, as in the traditional view of ocean heat transport ([Bryan, 1991]). Furthermore, the essential role played by mixing processes for balancing the heat budget is made very clear.

[4] Let us now consider the heat budget for the control volume that is bounded by the sea surface and the  $14^{\circ}\text{C}$  isentrope. The  $14^{\circ}\text{C}$  isentrope spans roughly the range of latitudes between about  $40^{\circ}\text{N}$  and  $40^{\circ}\text{S}$  (see Figure 1). We use the ocean heat transport estimates shown in the work of Wunsch [2005, Figure 3] [see also Ganachaud and Wunsch, 2000] to estimate the total surface heat input to our control volume to be about 1 PW (1 PW is  $10^{15}$  W). It should be noted that the error bars are such that the actual net heat input could be as much as 2 PW or as little as 0 PW (as implied by [Grist and Josey, 2003]). Taking  $\mathcal{H} = 1$  PW,  $K_s = 10^{-5} \text{ m}^2 \text{ s}^{-1}$ , as found from microstructure measurements ([Ledwell *et al.*, 1993]), and  $\partial \bar{T} / \partial n = 2^{\circ}\text{C} / 100 \text{ m}$  from Figure 1, the integrated contribution from the small-scale mixing is almost one order of magnitude too small to balance the surface heat input. Furthermore, since the  $14^{\circ}\text{C}$  isentrope is confined to the upper few hundred meters of the water column, we cannot invoke enhanced mixing over rough topography [e.g., Ledwell *et al.*, 2000] to close the budget. It is also easy to show that 3-D mixing in the surface mixed layer associated with the  $K_s$  term is insufficient to balance the budget, since the horizontal length scales associated with 3-D mixing are too small to provide the necessary diffusivity. It follows that if one accepts that the net heat input is 1 PW, then to close the heat budget one needs to invoke the  $K_e$  term; that is, mixing associated with the departures from the zonal/time average. Diagnoses using the  $16^{\circ}\text{C}$  or  $18^{\circ}\text{C}$  isentropes add support to this conclusion because as the water mass contained in our control volume warms, one can be increasingly confident that the lower bound on the estimate for the net heat input,  $\mathcal{H}$ , is significantly above 0 PW.

[5]  $K_e$  is the diffusivity arising from the departures from the zonal/time average and we can use the theory of [Eden *et al.*, 2007] to estimate its value. In our case, these departures are associated with mesoscale eddies, standing eddies arising from the departure of the time-mean flow from the zonal average, and also the seasonal cycle. The latter is a complication we expect to be associated with a negative diffusive effect. This is because when the variations in the thermal forcing,  $Q'$ , are seasonal only, the temperature and forcing changes,  $T'$ ,  $Q'$  respectively, are positively correlated, as can be seen in the work of Gill and Turner [1976, Figure 1], leading to a negative diffusivity



**Figure 2.** Surface eddy diffusivity in units of  $\text{m}^2 \text{s}^{-1}$  estimated from equation (11) and color coded using a log scale to base 10, together with the mean SST ( $^{\circ}\text{C}$ ) contours.

using equation (2). (Note that the effect of seasonal forcing is to enhance the equator to pole temperature contrast, rather than reduce it.) Clearly, however, if  $K_e$  is negative, then the heat budget given by equation (4) cannot be balanced. It follows that either the mesoscale eddies or the standing eddies dominate. A second complication is the role of along-isopycnal mixing. Since potential temperature is the dominant influence on density over most of the ocean (at least the subtropical regions we are interested in), we neglect this effect in what follows. On the other hand, the eddy-induced diffusivity, whether this comes from the mesoscale or the standing eddies, will be important in the surface mixed layer where the ocean has contact with the atmosphere [e.g., *Tandon and Garrett, 1996*]. We therefore assume that  $K_e$  is dominated by the contribution from the surface mixed layer and write equation (4) as

$$2\bar{h}K_eL_w \frac{\partial \bar{T}}{\partial y} \approx \frac{\mathcal{H}}{\rho c_p} \quad (5)$$

where we have neglected the contribution from 3-D mixing (the  $K_s$  term) and  $\bar{h}$  represents a mean mixed layer depth. To estimate  $\bar{h}$ , we use the climatology of the mixed layer depth taken from the U.S. Naval Research Laboratory ([*Kara et al., 2003*]), which is on the order of 100 m at  $40^{\circ}$  latitude in both hemispheres. The diffusivity,  $K_e$ , diagnosed from equation (4) using  $\mathcal{H} = 1$  PW is then close to  $10^4 \text{ m}^2 \text{ s}^{-1}$ .

[6] The eddy diffusivity in the surface mixed layer can also be estimated directly from theory. We neglect the higher order terms (which can be shown to be small) in equation (2) to obtain

$$K_e |\bar{T}_y|^2 \approx -\overline{Q'T'} \quad (6)$$

where  $\bar{T}_y$  is the meridional gradient of the zonally-averaged sea surface temperature (SST), and  $T'$  is the departure of the SST from the zonal/time average. To illustrate the importance of the damping of SST variance by the

atmosphere, we replace  $Q'$  by a simple restoring boundary condition, in which SST is relaxed back to climatology on a time scale  $(1/\gamma)$  of 50 days [*Haney, 1971*], so that

$$K_e |\bar{T}_y|^2 \approx \gamma \overline{T'^2}. \quad (7)$$

In order to estimate  $\overline{T'^2}$ , we use climatological SST data from the World Ocean Atlas and compute the variance based on the departures from the zonal average. Taking  $|\bar{T}_y|$  from Figure 1, then gives values of  $K_e$  that are significantly larger than  $10^4 \text{ m}^2 \text{ s}^{-1}$ , indicating that sufficient mixing is indeed available in the surface mixed layer to close the ocean heat budget in the zonally-averaged case. It should be noted that this estimate for the diffusivity is based on the standing eddies only (analogous to the gyre component in [*Bryan, 1962*]) and does not include the effect of mesoscale eddies. In the 3-D case discussed next, only transient eddies are available to provide the necessary mixing.

### 3. The 3-D Case

[7] Following [*Eden et al., 2007*], the equation for the ocean heat budget in the 3-D case in statistically steady state is:

$$\nabla \cdot (\mathbf{u}^* \bar{T}) = \bar{Q} + \nabla \cdot (K_e \nabla \bar{T}) \quad (8)$$

where here the overbar represents a time mean carried out in height coordinates,  $\nabla$  is now a 3-D operator,  $\mathbf{u}^*$  the 3-D “residual mean” velocity (different from  $\mathbf{u}^*$  in equation (1)) and  $K_e$  is the diffusivity in the 3-D case given in statistical steady conditions by

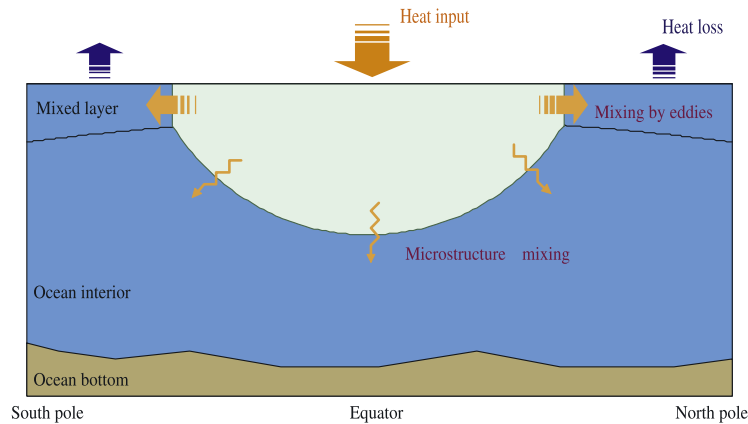
$$K_e |\nabla \bar{T}|^2 = -\overline{Q'T'} + \text{higher order terms}. \quad (9)$$

As before, the diffusivity,  $K_e$ , in general will be positive (in association with the irreversible removal of variance). Taking the control volume to be the volume of water between the sea surface and a mean isentropic surface, say  $T = T_0$ , it follows, exactly as before, that the residual advection term drops out from the heat balance, again revealing the central role played by mixing processes, and leaving the balance

$$\int_A (K_s + K_e) \frac{\partial \bar{T}}{\partial n} dA = \frac{\mathcal{H}}{\rho c_p} \quad (10)$$

where the integral is now taken over the isentropic surface  $T = T_0$ , and  $\mathcal{H}$  is the total surface heat input to the control volume. We now use the surface heat flux climatology of [*Grist and Josey, 2003*] to adjust our estimate for the net surface heat input in the zonally-averaged case in order to account for the fact that the  $14^{\circ}\text{C}$  isentrope does not outcrop exactly along the latitude lines  $40^{\circ}\text{N}$  and  $40^{\circ}\text{S}$ . This adjustment turns out to be negligibly small compared to 1 PW, and we therefore take 1 PW as our estimate for the net surface heat input to our control volume in the 3-D case, as for the zonally-averaged case (while recognising the uncertainty in this estimate noted above). It follows, as before, that we expect mixing associated with  $K_e$  to play an important role in balancing the ocean heat budget.





**Figure 3.** Schematic illustrating the oceanic heat budget.

Since in the 3-D case we are using time averaging, the influence of standing eddies is excluded, and mesoscale eddies are the only mechanism available to provide the mixing necessary to overwhelm the negative diffusive effect from the seasonal cycle. We now concentrate, as before, on the mixing arising due to interaction between mesoscale eddies and the surface heat flux. To provide some indication of the likely magnitude of  $K_e$  in this case, we replace  $Q'$  in equation (9) by a simple restoring boundary condition, as before, and neglect the higher order terms to obtain

$$K_e |\nabla \bar{T}|^2 \approx \gamma \bar{T}^2. \quad (11)$$

where  $\bar{T}^2$  is the SST variance which, here, is derived from satellite data. The SST data have a resolution of 14 km and are taken from the NOAA Satellite and Information Service website (details available at <http://www.class.noaa.gov/nsaa/products>). The selected dataset spans the period from August 2001 to September 2005, and is available once a week. The SST anomaly ( $T'$ ) is computed after the seasonal cycle has been removed and is the departure from the mean over the whole study period. Figure 2 shows the estimated diffusivity for the Gulf Stream region computed from the variance using equation (11) and a time scale of 50 days for  $1/\gamma$ . In the region of the Gulf Stream front, the estimated values are of order  $1000 \text{ m}^2 \text{ s}^{-1}$  due to the strong mean gradient in SST there. Much larger values (of order  $10^4 \text{ m}^2 \text{ s}^{-1}$ ) are found immediately south of the Gulf Stream where the mean gradient is relatively weak. We note that the net northward flux of heat computed by multiplying the local value of the diffusivity by the local value of the gradient of mean SST is comparable to the flux implied by equation (5), suggesting that as in the zonally-averaged case, there is sufficient mixing available in the surface mixed layer due to air-sea interaction processes alone to balance the 3-D ocean heat budget. Clearly future work should focus on estimating the diffusivity  $K_e$  globally and on developing parameterizations for  $K_e$  for use in climate models. A first attempt at estimating the diffusivity from satellite data has been given by *Zhai and Greatbatch* [2006a] and agrees quite well in both amplitude and spatial structure with the

estimate in Figure 2. *Zhai and Greatbatch* [2006b] have estimated the eddy-induced surface diffusivity from a model and also find a similar pattern and amplitude. In particular, the large values, approaching  $10^4 \text{ m}^2 \text{ s}^{-1}$  immediately south of the Gulf Stream, are feature of diagnoses from both observations [*Zhai and Greatbatch*, 2006a] and models [*Zhai and Greatbatch*, 2006b] and appear to be robust.

#### 4. Conclusions

[8] Equations (4) and (10) show that mixing (either 3-D mixing acting on the mean gradient or eddy-induced mixing) is the essential ingredient for closing the ocean heat budget. Our results have emphasised the diabatic aspect of the mesoscale eddies [*Tandon and Garrett*, 1996]. In particular, we argue that eddy-induced mixing in the surface mixed layer due to air-sea interaction processes can play an important role in closing the ocean heat budget (illustrated schematically in Figure 3). It follows that the lateral mixing applied near the surface in non-eddy resolving ocean/climate models may be required to play an important role in closing the ocean heat budget in these models, and that careful attention should be given to how the lateral mixing in these models is specified. The current practise is often to simply replace the isopycnal mixing in the ocean interior by horizontal mixing at the surface, with no guarantee that the magnitude and spatial structure of the mixing is appropriate. There is clearly a need to develop physically-based parameterizations for eddy-induced mixing, especially in the surface mixed layer. The important role played by the damping of SST variance by the surface heat flux has been emphasised in this paper, an effect that has been illustrated using a model by [*Zhai and Greatbatch*, 2006b].

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C. Eden, IFM-GEOMAR, Düsternbrooker Weg 20, D-24105 Kiel, Germany.

R. J. Greatbatch and X. Zhai, Department of Oceanography, Dalhousie University, Halifax, NS, Canada B3H 4J1. (richard.greatbatch@phys.ocean.dal.ca)

D. Olbers, Alfred Wegener Institute for Polar and Marine Research, Bussestrasse 24, D-27570 Bremerhaven, Germany.