

Water mass transformation due to mixed layer entrainment and mesoscale stirring: In series or parallel?

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Abstract. The convergence of advective and diffusive buoyancy flux must match the air-sea buoyancy flux between two outcropping isopycnals. This leads to a diagnostic framework for water mass transformation in which a myriad of different processes can be incorporated under a unifying balance. We review how the diapycnal advection due to ubiquitous mixed layer entrainment can be included in this framework, and we estimate its contribution to the large scale transformation. We also consider how decomposing the flow and buoyancy field into mean, eddy and turbulent parts leads to clarifying the interaction of mixed layer and mesoscale (or sub-mesoscale) eddies in the overall large scale balance.

1. Introduction

Quantifying the water mass changes and finding the water mass formation rates for specific density classes are central to understanding the diabatic ocean circulation. Accurate quantification of water mass formation rates is also important for understanding variability in oceanic heat and freshwater fluxes and their contribution to climate.

The current description of water mass transformation was formulated by *Walín* (1982), who proposed that the poleward surface drift in the ocean can be determined directly by the air-sea heat flux $-B_0$ at the surface.¹ This led Walín to suggest the relation

$$F(T) = B_0 \frac{dS}{dT} \quad (1)$$

where \mathbf{F} is the cross-isotherm advection (volumetric flow rate), and δS is the area enclosed between the two isotherm outcrops as shown in Figure 1.

This relationship was subsequently successfully used in diagnostic calculations (*Tziperman* 1986, *Speer and Tziperman* 1992). These calculations showed close correspondence between the water mass formation derived by the diathermal advection from the Walín relation and the estimated rates of water mass formation for

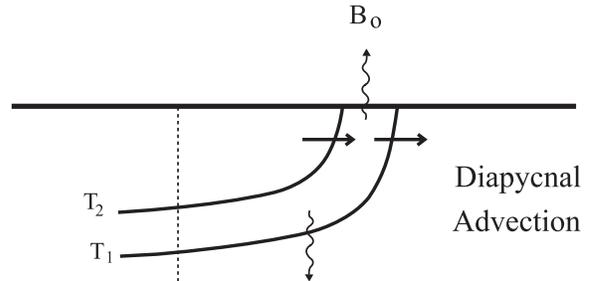


Figure 1. A schematic of the diapycnal advection $F(T)$ considered by *Walín* (1982). The net surface heat flux is balanced by a net advective flux at the surface.

Sub-Tropical mode water and the Sub-Polar Mode Water from observations. However, the *Walín* (1982) formulation assumed negligible mixing in the upper ocean.

In the same year that Walín proposed his ideas about this diathermal advection, *Niiler and Stevenson* (1982) sought to constrain the values of diapycnal mixing by considering the heat budget of the isotherms that outcrop in the tropics and are subject to net heat gain through the year as shown in Figure 2.

In this case, since by definition the net change in the mean annual isothermal position is zero, there cannot be a net diathermal advection and the net heat flux must be balanced by a net diffusive flux due to mixing. Therefore, at least in the limiting case of closed mean isopycnal surfaces, the role of diffusive flux due to mixing cannot be ignored, and the surface drift of

¹The interest in water mass formation rate in specific density classes has followed various diagnostics, e.g. in *Walín* (1982) arising out of considering budgets with heat fluxes and isotherms, and in other studies, the distribution of isopycnals and buoyancy fluxes. Thus we retain the symbol B_0 for both surface heat fluxes and surface buoyancy fluxes.

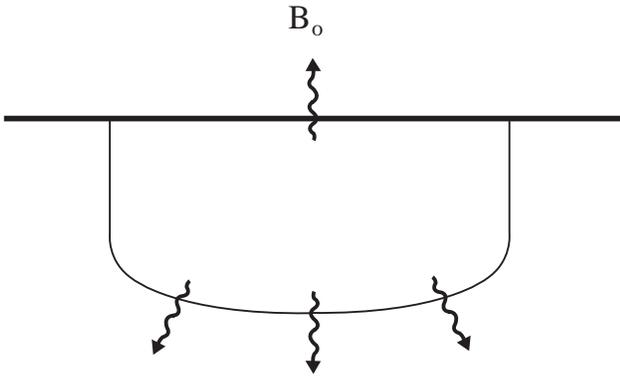


Figure 2. A schematic of the outcropping tropical annual mean isotherms considered by *Niiler and Stevenson (1982)*. The net surface heat flux must be balanced by a net diffusive flux due to mixing.

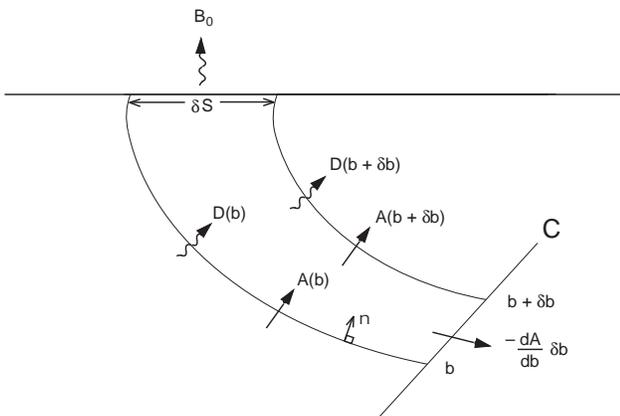


Figure 3. Air-sea buoyancy flux $B_0 \delta S$ between two outcropping isopycnals b and $b + \delta b$ must be balanced by convergence of diapycnal advective flux due to advection $A(b)$ and diffusive flux $D(b)$. From *Garrett et al. (1995)*.

Walín (1982) cannot achieve this balance.

This realization led *Garrett et al. (1995)* to clarify the *Walín* formulation. They insisted that considering the diapycnal advection about any averaged isopycnal position would require diapycnal mixing, since otherwise the isopycnal would simply be advected around by the flow. As *Figure 3* shows, the convergence of the diapycnal advective and diffusive fluxes must match the air-sea buoyancy flux between two isopycnals.

The *Walín (1982)* approach was thus generalized by

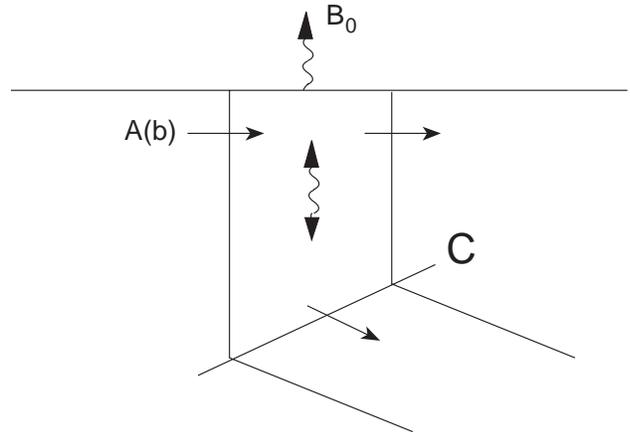


Figure 4. *Garrett et al. (1995)*'s ideal case. With no horizontal mixing and only vertical mixing, $D(b)$ can be ignored and the diapycnal advection equals the air-sea transformation, or, $A = F$.

Garrett et al. (1995) as

$$A(b) = F(b) - \frac{dD}{db} \quad (2)$$

where $D(b)$ is the diapycnal diffusive flux across the isopycnal b . Formal derivations of this relation appear in *Garrett et al. (1995)* and in *Marshall et al. (1999)*.

$D(b)$ consists of many mixing processes that can result in diffusive flux across isopycnal b . There will be contributions to D from diffusive mixing in the thermocline, while horizontal diffusive fluxes due to mesoscale stirring would arise near the ocean surface, and tidal mixing near rough bathymetry for the isopycnals that intersect with the ocean bottom in such regions. In so far as the balance in the upper ocean is concerned, if the control volume is chosen to be bounded by two isopycnals b and $b + \delta b$ and a control surface at the bottom of the mixed layer (*Figure 4*), *Garrett et al. (1995)* showed that for negligible horizontal mixing and with vertical isopycnals, vertical mixing $D(b)$ makes no contribution and the net diapycnal advection is only due to the air-sea transformation, that is,

$$A(b) = F(b) \quad (3)$$

where $F(b) = -B_0 \frac{dS}{db}$ is the air-sea transformation, with the negative sign due to the convention that the diapycnal advection from less buoyant to more buoyant water mass is considered positive (reversing the signs of *Walín's* convention in *Figure 1*). This is an important idealization, since in this case the diapycnal advection can be fully diagnosed by surface buoyancy flux and surface buoyancy outcrop distribution.

Speer (1997) evaluated $F(b)$ and $A(b)$ using a hydrographic section at 11°S , as shown in the schematic

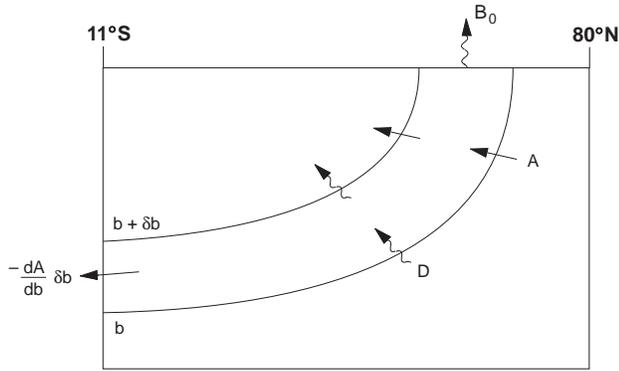


Figure 5. A schematic showing the control volume for the diagnostic calculation carried out by *Speer (1997)*.

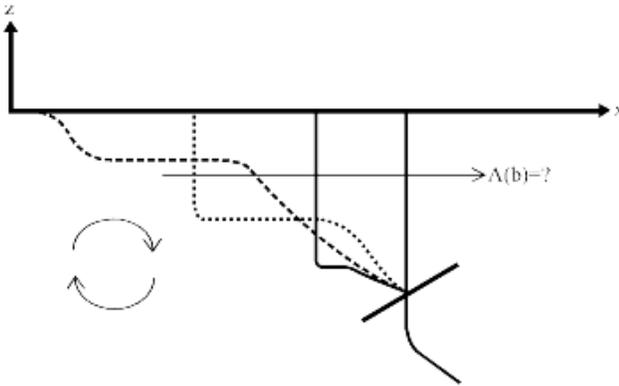


Figure 6. The ideal algorithm can no longer be expected to be valid if time-dependent effects are taken into account with entrainment at the mixed layer base.

Figure 5. These diagnostic calculations showed that the residual between $A(b)$ and $F(b)$ is significant, and therefore transformation due to the diffusive flux term $dD(b)/db$ must be considered. While this diagnostic calculation cannot be used to pinpoint which processes contribute to the D term, *Speer (1997)* showed that the calculations are consistent with diapycnal mixing in the thermocline.

The formulation considered thus far has assumed steady state, but an accumulation within the isopycnals can be interpreted as a net inflow, and (2) can still be used. A steady state scenario neglects the role of diapycnal advection due to time dependent deepening and shallowing of the mixed layer, at both seasonal and diurnal timescales. Conceptual modifications are required to relax the steady state assumption for the water mass transformation balance as shown in Figure 6.

The next section considers diapycnal advection due to the ubiquitous process of mixed layer deepening and entrainment. Upwelling at the base of the mixed layer

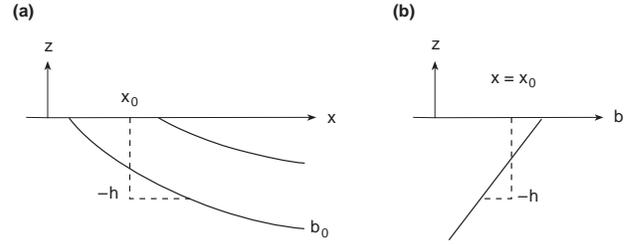


Figure 7. A schematic showing isopycnal distribution (left) and buoyancy distribution (right). The solid lines are before the mixing event and the dashed lines are after the mixing event.

and subsequent mixing results in water mass transformation which is conceptually very similar to that due to mixed layer deepening and entrainment at the base of the mixed layer. Regions of significant upwelling such as those found near submesoscale instabilities near fronts are considered next via an example from FASINEX (*Weller 1991*). The final section considers whether the stirring due to mesoscale processes should be considered as an independent process for water mass formation and hence considered to be independent of the transformation due to the mixed layer entrainment, or, do the two processes happen together, in series and are interdependent. The water mass transformation rates would only be additive in the former case. We conclude with a summary and unanswered questions.

2. Transformation Due to Mixed Layer Deepening

Garrett and Tandon (1997) show that buoyancy redistribution in the vertical by mixing achieves diapycnal advection across an isopycnal surface. Figure 7 shows an isopycnal that outcrops at position x_0 due to vertical mixing, and the buoyancy balance is shown on the right. x is taken to be the direction normal to the outcrop towards less dense waters.

With a uniform buoyancy gradient in the vertical and in the horizontal everywhere, the volume between the successive isopycnals does not change, i.e, there is no diapycnal advection. However, a departure from these conditions implies a net diapycnal advection. In particular, if a pre-existing mixed layer of depth $-h + dh$ deepens to $-h$, as shown in Figure 8, *Garrett and Tandon (1997)* show that the net diapycnal advection F_{ML} is given by

$$F_{ML} = \frac{1}{\mathcal{T}} \int_0^{\mathcal{T}} \left[\left(\frac{1}{b_{xi}} - \frac{1}{b_{xs}} \right) \Delta b - \frac{1}{2} b_{xx} b_x^{-3} (\Delta b)^2 \right] \left(w + \frac{\partial h}{\partial t} \right) dt \quad (4)$$

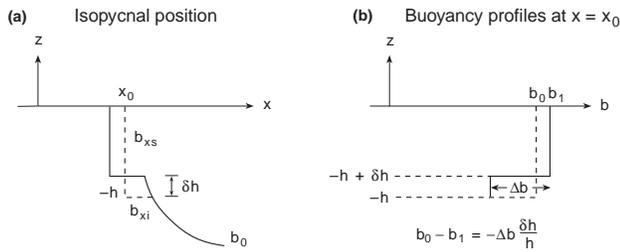


Figure 8. A schematic showing isopycnal distribution (left) and buoyancy distribution (right) for a pre-existing mixed layer in presence of both b_{xx} and b_{xz} . The solid lines are before the mixing event and the dashed lines are after the mixing event.

per unit outcrop length, the subscripts i and s refer to the gradients below the mixed layer (interior) and surface mixed layer respectively, Δb is the buoyancy change at the mixed layer base before a deepening event, h is the mixed layer depth, w is the vertical velocity at the mixed layer base, b_{xx}, b_x are buoyancy derivatives calculated at the surface, t is the time and $\mathcal{T} = 1$ yr. F_{ML} is the integrated diapycnal advection over the year for each deepening event $w + \frac{\partial h}{\partial t} > 0$, since the upwelling with vertical velocity w has an effect similar to mixing and subsequent deepening by $w \delta t$.

The F_{ML} expression above consists of two terms, the first depending on vertical variation of horizontal buoyancy gradients and the second depending on the curvature or the horizontal variation of the surface buoyancy gradients. *Garrett and Tandon (1997)* remark on several cases in the limit as b_{xi} , the buoyancy gradient at the mixed layer interface and b_{xs} , the buoyancy gradient at the surface vanish, showing that the singularity in (4) is no worse than the singularity in $F_{air-sea} = -B_0(dS/db)$. The formulae need information on both large-scale horizontal gradients and vertical cycling of the mixed layer.

Tandon and Zahariev (2001) have used the Marine Light Mixed Layer (MLML) experiment mooring observations (*Plueddemann et al. 1995*) which include both spring and fall mixed layer transitions and used a combination of mixed layer model results and hydrography (*daSilva et al. 1994*) to calculate the b_{xx} terms to get an order of magnitude estimate of F_{ML} . The sensitivity to averaging of synoptic events is also explored. Their calculations indicate that if hourly winds are used, the water mass transformation due to mixed layer entrainment has annual peak contributions of about $O(4)$ Sv for $\sigma_t = 24.0$. This is comparable to the annual transformation attained by diapycnal mixing in the upper ocean water masses by *Zhang and Talley (1998)*. However, with daily averaged winds and without diurnal

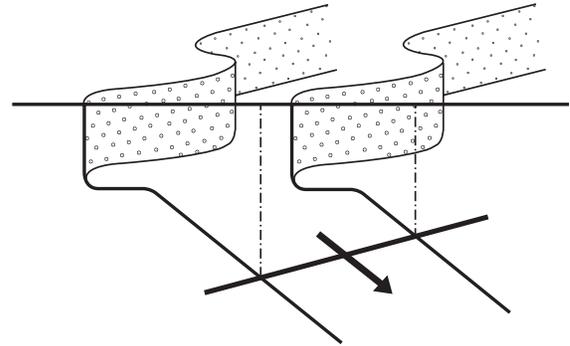


Figure 9. With meanders superposed on a mean isopycnal distribution, F_{ML} estimates are higher, both due to the variation in buoyancy gradient and the increased length of the isopycnal over which water masses get transformed.

variation in buoyancy forcing, this contribution is up to an order of magnitude smaller. Another set of mixed layer simulations includes an annual cycle with a shallow and strong summer thermocline. Inclusion of synoptic summer forcing for this scenario leads to transformation values several times larger than above, about $O(14)$ Sv at $\sigma_t = 24.0$. The peak contribution in this case is almost two orders of magnitude smaller if the synoptic forcing is averaged daily and the diurnal cycle is not resolved. The upper bound for the b_{xz} term at the MLML experiment site is $O(3)$ Sv. These estimates are likely to be underestimates since hydrographic data average over isopycnal meanders, due to both increase in the curvature terms and an increase in the isopycnal length (Figure 9). More details are presented in the appendix of *Tandon and Zahariev (2001)*. Contribution due to mesoscale and sub-mesoscale meanders is also conceptually considered later in more detail in section 4.2 of this paper.

3. Transformation Due to Upwelling Near Fronts

While the b_{xz} term was found to be small in *Tandon and Zahariev (2001)* away from the fronts, this need not be the case near fronts. Mesoscale eddies (wavelength ~ 500 km) strain the surface into $O(100)$ km tongues, which leads to sub-mesoscale features of 20-50 km diameter which induce vertical velocities of $O(40)$ m/day thereafter (e.g. *Pollard and Regier (1992)* for Frontal Air-Sea Interaction Experiment site, *Rudnick (1996)* and *Rudnick and Luyten (1996)* for the Azores front).

Due to the rapid upwelling near the fronts, the upwelling at the base of the mixed layer $\int_w dt_{>0} w dt$ dominates the deepening $\int_{dh>0} dh$ in the F_{ML} term. Using typical FASINEX values of $b_{xs} = 1.5 \times 10^{-7} \text{ s}^{-2}$ (for temperature gradient of 1K/20 km), $b_{xx} = 1.5 \times 10^{-12} \text{ m}^{-1} \text{ s}^{-2}$ and $\Delta b = 10^{-3} \text{ m s}^{-2}$. With upwelling of O(40 m/day) along half the length of a convoluted frontal length of 3000 km, the annualized transformation values are O(0.2 Sv) for the b_{xx} term and O(5 Sv) for the b_{xz} term. The F_{ML} term is therefore likely to have a greater contribution from b_{xz} term and upwelling near sub-mesoscale features.

To summarize, while F_{ML} estimates are O(5)Sv, they are particularly sensitive to the resolution of synoptic events and sub-surface density structure near fronts. It is also reasonable to ask whether these effects are already assumed to be present in a horizontal diffusive flux due to mesoscale stirring. Inspired by *Garrett's* review (this volume) as well as *Nakamura* (1996) and *Winters and D'Asaro* (1996), we next ask, under what circumstance can these two processes result in water mass transformation independent of each other and when must they be considered interdependent.

4. Mesoscale Stirring and Mesoscale-Mixed Layer Interaction

Horizontal stirring due to mesoscale eddies and subsequent mixing would contribute to $D(b)$, and sub-mesoscale upwelling contributes to $F_{ML}(b)$. How should the distinction be made between water mass transformation due to these two processes? How do we ensure that the transformation isn't being diagnosed multiple times under different processes which are in series (and hence not independent)? These questions are clearly related to the how the resolved and unresolved processes are defined.

Recent understanding (*Speer et al.* 1997; *Garrett and Tandon* 1997; *Marshall et al.* 1999, *Nurser et al.* 1999) has concentrated on expanding (2) heuristically in terms of the physical processes, which can be written as

$$A(b) = A_{\text{Ekman}}(b) + A_{\text{eddy}}(b) + A_{\text{mean}}(b) \quad (5)$$

$$= F_{\text{air-sea}} + F_{ML} - \frac{dD_{\text{eddy}}}{db} - \frac{dD_{\text{int}}}{db} \quad (6)$$

with (5) for the dynamics which must balance (6) for the thermodynamics. The dynamical processes are the diapycnal volume flux due to wind forcing (Ekman), mesoscale diapycnal volume flux, and the mean geostrophic flow across isopycnals. The thermodynamic processes are the air-sea transformation $F_{\text{air-sea}}$, the contribution due to mixed layer entrainment F_{ML} , horizontal mixing due to eddies dD_{eddy}/db , and the interior

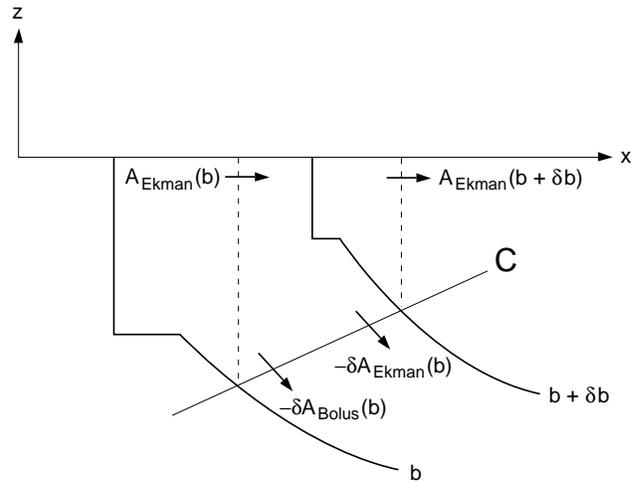


Figure 10. An illustration showing the modified control volume. $A_{\text{Ekman}}(b)$ refers to the wind-driven diapycnal volume flux past the control volume surface b , while $A_{\text{Bolus}}(b)$ or $A_{\text{eddy}}(b)$ refers to the diapycnal volume flux due to the mesoscale eddies past this control surface. The lower control surface C is the base of the winter mixed layer. Dashed vertical lines show the lateral control surfaces for the dynamic method; the instantaneous isopycnal positions shown as solid lines form the lateral control surfaces for the thermodynamic method.

diapycnal mixing dD_{int}/db . This procedure is naturally isopycnic/diapycnic and allows comparison of these diverse physical processes under a single framework.

The terms that are most readily calculated from climatological data in the above framework are A_{Ekman} and $F_{\text{air-sea}}$, e.g. the recent evaluations of A_{Ekman} and $F_{\text{air-sea}}$ for the Southern Ocean by *Speer et al.* (1997). The evaluation of A_{Ekman} is generally done following isopycnal surfaces around the annual cycle. However, the control volumes for the dynamic and thermodynamic methods can be chosen differently as long as they coincide at the winter mixed layer depth. Thus, a modified control volume for the dynamic calculation can be chosen (*Garrett and Tandon* 1997) such that this consists of the volume between the vertical projections of the isopycnal surfaces below the winter mixed layer to the surface (Figure 10). Therefore, this modified controlled volume does not change much during the annual cycle.

4.1. The triple decomposition into mean, eddy, and turbulent flow

While choosing a fixed control volume as described above simplifies the diagnostic evaluations of the dynamic terms in (5), the issues concerning the decompo-

sition and overlap of the thermodynamic terms in (6) still remain. To clarify the role of horizontal stirring dD_H/db and mixed layer entrainment F_{ML} , following *Davis* (1994) and *Garrett* (this volume), we apply the triple decomposition to the water mass transformation ideas. As these authors stress, such a decomposition necessarily presupposes a gap in time (or space) scales such that a three-way decomposition can be defined. For the water mass transformation diagnostics the buoyancy field is split into the mean, eddy, and turbulent parts. The diapycnal advection u_n is also split into three parts, each representing the mean, eddy, and turbulent advection (subscripts m, e , and t respectively) across the mean buoyancy surface, i.e.,

$$\begin{aligned} b &= b_m + b_e + b_t & \text{and} & \quad (7) \\ u_n(b_m) &= u_{nm}(b_m) + u_{ne}(b_m) + u_{nt}(b_m). & (8) \end{aligned}$$

Here we envisage the mean buoyancy field b_m to be the modified mean buoyancy field (*Killworth* 2001, *McDougall and McIntosh* 2001) although for the large scale water mass transformation diagnostics the difference with Eulerian mean buoyancy field is probably small. We denote $\overline{(\)}$ for averaging over a time scale large compared to turbulence but small compared to that of the eddies and $\langle (\) \rangle$ for averaging over a time scale large compared to that of the eddies but short compared to that of the mean. Substituting into the instantaneous buoyancy equation and subsequent averaging over the turbulent time-scales and the eddy time scale yields

$$u_{nm} \cdot \frac{\partial b_m}{\partial n} + \langle u_{ne} \cdot \frac{\partial b_e}{\partial n} \rangle = - \langle \overline{u_{nt} \cdot \frac{\partial b_t}{\partial n}} \rangle \quad (9)$$

if the mean flow is considered steady at long time-scales. The right hand side of eq.(9) is the $-\frac{\partial D}{\partial n}$ term and the terms on the left hand side are mean and eddy advection of mean and eddy isopycnals respectively. This equation is subject to the surface boundary condition

$$\langle w_e b_e \rangle + \langle \overline{w_t b_t} \rangle = B_0 \quad (10)$$

thus suggesting that the surface buoyancy flux is balanced by both the vertical buoyancy flux due to eddies and turbulent flux at the surface. *Garrett* (this volume) discusses the buoyancy variance budget for this decomposition.

4.2. Mesoscale transformation and the entrainment contribution: In series or parallel?

If the diagnostics are done in a framework that does not resolve eddies, then the transformation relationship can be written in terms of buoyancy balance about b_m :

$$A(b_m) = A_{\text{mean}}(b_m) + A_{\text{eddy}}(b_m) + A_{\text{Ekman}}(b_m) \quad (11)$$

$$= F(b_m) - \frac{dD_H}{db}(b_m) - F_{ML}(b_m) - \frac{dD_v}{db}(b_m) \quad (12)$$

where the contribution $\frac{dD_H}{db}(b_m)$ is due to the convergence of the eddy stirring horizontal diffusive flux and $F_{ML}(b_m)$ is the contribution to water mass transformation due to mixed layer deepening and entrainment, both evaluated about the mean isopycnal positions.

Following the triple decomposition suggested above, if the eddies are resolved in this framework, the water mass transformation balance becomes

$$\begin{aligned} A(b_m) &= A_{\text{mean}}(b_m) + A_{\text{eddy}}(b_m) + A_{\text{Ekman}}(b_m) \quad (13) \\ &= F(b_m + b_e) - F_{ML}(b_m + b_e) - \frac{dD_v}{db}(b_m + b_e) \quad (14) \end{aligned}$$

and the eddy-stirring component $\frac{dD_H}{db}$ does not arise. The contributions in eq. (11) and eq. (13) are the same due to the choice of control volume in Figure 10. However, the contributions in eq. (12) and eq. (14) are different. In this case, the diapycnal advection due to horizontal mixing by the eddies and the contribution due to mixed layer entrainment are in *parallel* if the budgets are considered about the mean buoyancy surfaces (b_m) that do not resolve the eddies as in eq. (12). However, these two processes are in *series* if budgets are considered for eddy buoyancy surfaces ($b_m + b_e$) as in eq. (14), where eddy isopycnals are defined by modified mean density that averages over a time scale long compared to the turbulence but short compared to the eddies. Thus the F_{ML} estimates computed by *Tandon and Zahariev* (2001) (section 2) which are based on mean monthly isopycnals that average over the eddies correspond to the estimates for $F_{ML}(b_m)$, while the estimates earlier near frontal zones (section 3) correspond to the latter case $F_{ML}(b_m + b_e)$.

5. Discussion

The uncertainty in F_{ML} remains unacceptably high in both model based estimates of coarse non-eddy resolving simulations (e.g. *Nurser et al.* 1999) and data based evaluations which are based on localized mixed layer simulations (*Tandon and Zahariev* 2001). More accurate basin wide estimates are necessary to make progress. While considering budgets in an eddy resolving framework, the water mass transformation due to eddies happens *in series* with diapycnal mixing and entrainment. In this case, another significant concern is identifying the rate controlling process for diapycnal transformation in the upper ocean—do the mesoscale eddies define the rate at which net diapycnal advection will happen or does it primarily depend on the rate at

which diapycnal mixing occurs? The mesoscale stirring would be dominant for water mass transformation if the mixing processes are so efficient in the upper ocean that they act to dissipate the eddies quickly and transform water masses on a fast time scale. In this case, it is the baroclinic instability and larger scales that would set the net diapycnal advection rates. On the other hand it is possible that the mixed layer processes allow non-linear interactions amongst eddies, and the net diapycnal advection is then significantly dependent on mixing processes in the upper ocean. While recent eddy resolving simulations have concentrated on adiabatic properties of mesoscale parameterizations (e.g., articles by Marshall, Killworth, and McDougall from this volume) diabatic interactions in the upper ocean remain unexplored, and the question above remains unanswered. Determining which process is dominant for water mass transformation rate should be feasible with multiple eddy-resolving simulations whose sensitivity to mixing in the upper ocean and air-sea interaction is explored systematically.

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References

- daSilva, A. M., C. C. Young, and S. Levitus, Atlas of surface marine data 1994, vol. 1: Algorithms and procedures, U.S. Department of Commerce, NOAA, NESDIS, 1994.
- Davis, R. E., Diapycnal Mixing in the Ocean: The Osborn-Cox model, *J. Phys. Oceanogr.*, *24*, 2560–2576, 1994.
- Garrett, C., and A. Tandon, The effects on water mass formation of surface mixed layer time-dependence and entrainment fluxes, *Deep-Sea Res.*, *44/12*, 1991–2006, 1997.
- Garrett, C., K. Speer, and E. Tragou, The relationship between water mass formation and circulation, with application to Phillips' Red Sea model, *J. Phys. Oceanogr.*, *25*, 1696–1705, 1995.
- Marshall, J., D. Jamous, and J. Nilsson, Reconciling thermodynamic and dynamic methods of computation of water-mass transformation rates, *Deep-Sea Res.*, *46*, 545–572, 1999.
- Niiler, P. P., and J. Stevenson, On the heat budget of tropical warm pools, *J. Mar. Res.*, *40(Suppl.)*, 465–480, 1982.
- Nurser, A. J. G., R. Marsh, and R. G. Williams, Diagnosing water mass formation from air-sea fluxes and surface mixing, *J. Phys. Oceanogr.*, *29*, 1468–1487, 1999.
- Plueddemann, A. J., R. A. Weller, M. Stramska, T. D. Dickey, and J. Marra, Vertical structure of the upper ocean during the Marine Light-Mixed Layers experiment, *J. Geophys. Res.*, *100*, 6605–6620, 1995.
- Pollard, R. T., and L. A. Regier, Vorticity and vertical circulation at an ocean front, *J. Phys. Oceanogr.*, *22*, 609–625, 1992.
- Rudnick, D. L., Intensive surveys of the Azores Front., 2, Inferring the geostrophic and vertical velocity fields, *J. Geophys. Res.*, *101*, 16,291–16,303, 1996.
- Rudnick, D. L., and J. R. Luyten, Intensive surveys of the Azores Front, 1, Tracers and dynamics, *J. Geophys. Res.*, *101*, 923–939, 1996.
- Speer, K., and E. Tziperman, Rates of water mass formation in the North Atlantic Ocean, *J. Phys. Oceanogr.*, *22*, 94–104, 1991.
- Speer, K., S. Rintoul, and B. Sloyan, Sub-antarctic Mode Water Formation by Air-Sea Fluxes, *International WOCE Newsletter*, December, 29–31, 1997.
- Speer, K. G., A note on average cross-isopycnal mixing in the North Atlantic ocean, *Deep Sea Res.*, *44/12*, 1981–1990, 1997/.
- Tandon, A., and K. Zahariev, Quantifying the role of mixed layer entrainment for water mass transformation in the north atlantic, *J. Phys. Oceanogr.*, *31*, 1120–1131, 2001.
- Tziperman, E., On the role of interior mixing and air-sea fluxes in determining the stratification and circulation of the oceans, *J. Phys. Oceanogr.*, *16*, 680–693, 1986.
- Walín, G., On the relation between sea-surface heat flow and thermal circulation in the ocean, *Tellus*, *34*, 187–195, 1982.
- Weller, R. A., Overview of the Frontal Air-Sea Interaction Experiment (FASINEX): A study of air-sea interaction in a region of strong oceanic gradients, *J. Geophys. Res.*, *96*, 8501–8516, 1991.
- Zhang, H.-M., and L. Talley, Heat and buoyancy budgets and mixing rates in the upper thermocline of the Indian and global oceans, *J. Phys. Oceanogr.*, *28*, 1961–1978, 1998.

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