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Estimates of Cabbeling in the Global Ocean

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ABSTRACT

Owing to the larger thermal expansion coefficient at higher temperatures, more buoyancy is put into the ocean by heating than is removed by cooling at low temperatures. The authors show that, even with globally balanced thermal and haline surface forcing at the ocean surface, there is a negative density flux and hence a positive buoyancy flux. As shown by McDougall and Garrett, this must be compensated by interior densification on mixing due to the nonlinearity of the equation of state (cabbeling). Three issues that arise from this are addressed: the estimation of the annual input of density forcing, the effects of the seasonal cycle, and the total cabbeling potential of the ocean upon complete mixing. The annual expansion through surface density forcing in a steady-state ocean driven by balanced evaporation–precipitation–runoff (E–P–R) and net radiative budget at the surface $Q_{\text{net}}$ is estimated as 74 000 m$^3$ s$^{-1}$ (0.07 Sv; 1 Sv $= 10^6$ m$^3$ s$^{-1}$), which would be equivalent to a sea level rise of 6.3 mm yr$^{-1}$. This is equivalent to approximately 3 times the estimated rate of sea level rise or 450% of the average Mississippi River discharge. When seasonal variations are included, this density forcing increases by 35% relative to the time-mean case to 101 000 m$^3$ s$^{-1}$ (0.1 Sv). Likely bounds are established on these numbers by using different $Q_{\text{net}}$ and E–P–R datasets and the estimates are found to be robust to a factor of $2^c$.

These values compare well with the cabbeling-induced contraction inferred from independent thermal dissipation rate estimates. The potential sea level decrease upon complete vertical mixing of the ocean is estimated as 230 mm. When horizontal mixing is included, the sea level drop is estimated as 300 mm.

1. Introduction

In an ocean with a heat and freshwater budget that is steady in time, but spatially variable, there is an apparent negative density flux due to the nonlinearity of the equation of state. This would imply that the ocean should expand with zero net forcing at the surface. That is, the ocean is generally heated in low latitudes where surface temperatures (and thermal expansion coefficients) are high and cooled in high latitudes where surface temperatures (and expansion coefficients) are low. The apparent thermal density flux is negative even though the global heat budget may be balanced, and would thus add buoyancy at the ocean surface. How can this be? This question was posed by Eric B. Kraus during the 19th International Liège Colloquium on Ocean Hydrodynamics in 1987 and answered by Chris Garrett and Trevor McDougall, who were in the audience. A detailed answer was given a few years thereafter by McDougall and Garrett (1992, p. 1956):

Does the ocean continually expand as a result of the nonzero density flux? No. Rather, “densification upon mixing” in the interior of the ocean can ensure that its average density does not change.

Further elaborations were given by Davis (1994) and Zahariev and Garrett (1997), particularly in regards to the apparent nonzero density flux due to the seasonal cycle.

While both temperature and salinity contribute to cabbeling, the nonlinear contraction due to mixing of salinity is at least one order of magnitude smaller than thermal contraction (McDougall and Garrett 1992). To first order, one may thus approximate the divergence equation [McDougall and Garrett (1992), Eq. (23)] as

$$ \mathbf{V} \cdot \mathbf{u} = -\frac{1}{2} \frac{\partial \alpha}{\partial \theta} \chi_T, \quad (1) $$

where $\mathbf{V} \cdot \mathbf{u}$ is the divergence of the velocity field, $\theta$ is potential temperature, $\alpha$ is the thermal expansion coefficient $\alpha = -(1/\rho)(\partial \rho/\partial \theta)|_{S.P}$, $\chi_T$ is the dissipation...
rate of the thermal variance, and $\rho$ is potential density. Overbars in all equations denote temporal averages, and primes denote fluctuating components, so that $\mathbf{u} = \mathbf{u} + \mathbf{u}'$. For typical values of $\partial\alpha/\partial\theta$ and the thermal dissipation rate $\chi_T$, McDougall and Garrett estimated the rhs term of Eq. (1) to be on the order of $10^{-13}$ s$^{-1}$. While this makes it negligible for many practical applications of measuring or modeling the ocean interior, it still has potential important consequences (McDougall and Garrett 1992; Davis 1994), particularly for water mass transformations (Iudicone et al. 2008) and advection and diffusion (Klocker and McDougall 2010). Some further consequences are also highlighted in our results and conclusions.

The conundrum posed by E. Kraus is readily answered by the global integral of Eq. (1). The apparent expansion induced by the negative density flux (and hence buoyancy input) at the surface is compensated by the contraction from the destruction of thermal variance (mixing) in the interior of the ocean. The positive relation between surface heat fluxes and the thermal expansion coefficient generates a net expansion that is compensated by contraction associated with the interior mixing [Eq. (1)]. Here we evaluate the magnitude of the global net density flux imbalance and its implications for ocean cabbeling. We also relate these results to estimates of the global thermal dissipation rate after Joyce (1980).

The mean density equation at the surface and in steady state (meaning that $\partial\bar{\rho}/\partial t = 0$ and $\bar{\mathbf{u}} \cdot \nabla \bar{\rho} = 0$) may, to a first-order approximation, be adapted from McDougall and Garrett [1992, Eqs. (23) and (29)] to be written as

$$
\int_V \nabla \cdot (\bar{\rho} \mathbf{u}) \, dV = -\rho \int_V \nabla \cdot \mathbf{u} \, dV = \frac{1}{2} \int_V \left( \frac{\partial \alpha}{\partial \theta} \chi_T + \frac{\partial \beta}{\partial S} S \right) \, dV,
$$

where $\beta$ is the haline coefficient of seawater $\beta = -(1/\rho)(\partial \rho/\partial S)|_{\mathbf{T}, \mathbf{P}}$, $S$ is sea surface salinity, and $\chi_S$ is the dissipation rate of salinity variance, which is analogous to $\chi_T$ for temperature. Because the variation of $\beta$ with $S$ is small, only the thermal component will be considered here. Using the Gauss divergence theorem, the first-order estimate of thermal component of the density flux may then be written as

$$
\int_A (c_p^{-1} \alpha Q_{\text{net}}) \, dA = -\rho \int_V \nabla \cdot \mathbf{u} \, dV = \frac{1}{2} \int_V \left( \frac{\partial \alpha}{\partial \theta} \chi_T \right) \, dV,
$$

where $c_p$ denotes the specific heat capacity of water, which is taken as constant for thermodynamic consistency, and $Q_{\text{net}}$ is the net radiative budget at the surface, defined as the sum of shortwave radiation $Q_{\text{SW}}$, longwave radiation $Q_{\text{LW}}$, sensible heat flux $Q_{\text{SH}}$, and latent heat flux $Q_{\text{LH}}$. This equation shows how the global area integral of surface density flux is balanced by the volume integral of the contraction due to mixing.

With the recent availability of satellite-derived datasets of global ocean freshwater forcing and global oceanic heat budgets (Schanze et al. 2010), it is now possible to make estimates of the total rate of the density flux (and thus the buoyancy flux, which is proportional but opposite to the density flux) into the global ocean. From such flux data, the cabbeling rate in the interior can be calculated if a steady state is assumed. The surface density flux $F_\rho$ at a given position may be written as

$$
F_\rho = -c_p^{-1} \alpha Q_{\text{net}} + \rho \beta S (E - P - R) = F_{\rho T} + F_{\rho S}, \quad (4)
$$

where $E$ is the evaporation rate, $P$ is the precipitation rate, and $R$ is the riverine freshwater flux. Because this equation is valid only at the ocean surface, changes in pressure and the resulting pressure term are negligible and thus ignored here. It is clear from Eq. (4) that the lhs term in Eq. (3) is just the thermal component of the density flux $F_{\rho T}$.

The nonlinearities in this equation stem from the changes of $\alpha$ and $\beta$, where the change of $\alpha$ with $\theta$ is at least one order of magnitude greater than the salinity term (McDougall and Garrett 1992). To illustrate this effect, we have plotted the relationship of $\alpha$ as well as its first derivative with respect to potential temperature against potential temperature in Fig. 1. The zonal distribution of sea surface temperature (SST), $\alpha$, $Q_{\text{net}}$, and $F_{\rho T}$ are shown in Fig. 2. The global distribution of density flux into the ocean is shown in Fig. 3. The most obvious areas of densification are the western boundary currents, particularly the Gulf Stream and the Kuroshio. In these currents, warm water is advected poleward, which leads to evaporation and sensible heat loss. As a result of this, the surface water is cooled, leading to densification.

2. Datasets

New estimates of ocean freshwater and heat fluxes have been compiled—consisting of $E$ from the Objectively Analyzed Air–Sea Fluxes (OAFlux) (Yu and Weller 2007; Yu et al. 2008), $P$ from the Global Precipitation Climatology Project (GPCP) (Adler et al. 2003), and $R$ from Dai and Trenberth (2002) and Dai et al. (2009), which is presented in Schanze et al. (2010)—and used here for the freshwater forcing component $E-P-R$ and, hence,
The heat forcing component $Q_{\text{net}}$ and, hence, $F_{pT}$ is taken directly from the OAFlux product (Yu et al. 2008). We chose this product, because it matches the freshwater fields, and satellite-derived products have consistently outperformed reanalyses in in situ comparisons (e.g., Lorenz and Kunstmann 2012). Because we acknowledge that there is currently not a single $Q_{\text{net}}$ product that is clearly superior to others, we rerun all calculations with eight different $Q_{\text{net}}$ and four different $E-P-R$ products in section 3c. The temporal mean of the period 1987–2004 is taken for both $Q_{\text{net}}$ and $E-P-R$, as this is the common homogeneous period of available satellite and riverine runoff data (Schanze et al. 2010).

While the global ocean freshwater budget closes within the published errors of the individual products with a net imbalance of 0.5 Sv ($Sv = 10^6 \text{ m}^3 \text{ s}^{-1}$), the OAFlux heat budget has a net mean heat input to the ocean of $\sim 30 \text{ W m}^{-2}$. This exceeds estimates of actual global ocean warming of approximately 1 W m$^{-2}$ (Hansen et al. 2005).

For the buoyancy flux to be exactly compensated by cabbeling in the interior ocean, both $Q_{\text{net}}$ and $E-P-R$ must globally integrate to zero. By subtracting a constant value, we thus artificially balance both the global freshwater and the heat budget to be zero when integrated over the global ocean to fulfill the conditions

\begin{equation}
\iint_A (E - P - R) \, dA = 0,
\end{equation}

\begin{equation}
\iint_A (Q_{\text{net}}) \, dA = 0,
\end{equation}

where $Q_{\text{net}}$ is defined as the sum of shortwave radiation, longwave radiation, sensible heat flux, and latent heat flux.

The remaining required variables $S$ and $T$ are taken from the Met Office EN3 climatology product (Ingleby and Huddleston 2007) and are averaged over the same period of 1987–2004. The zonal averages of $Q_{\text{net}}, SST, \alpha$, and the thermal component of the buoyancy flux $F_{pT}$ are shown in Fig. 2. While there is a net addition of heat to the ocean near the equator, the main heat loss is in temperate latitudes and areas of western boundary currents. The warming at the equator gives rise to a negative density flux (and thus a positive buoyancy flux, caused

\begin{equation}
\iint_A (E - P - R) \, dA = 0,
\end{equation}

\begin{equation}
\iint_A (Q_{\text{net}}) \, dA = 0,
\end{equation}

where $Q_{\text{net}}$ is defined as the sum of shortwave radiation, longwave radiation, sensible heat flux, and latent heat flux.
by thermal expansion) while the cooling in temperate latitudes leads to densification.

While we consider the aforementioned datasets our “baseline,” the uncertainties in the covariances and error covariances between the products used in this study may result in a flawed estimate of the total error. We thus choose to use different commonly available data sets to indicate likely upper and lower bounds on these estimates. These products are described in more detail in section 3c.

Not all \( Q_{\text{net}} \) and \( E-P-R \) products have the same spatial (and temporal) coverage of the ocean owing to slightly different landmasks and different definitions of sea ice at high latitudes. To allow an efficient computation of the values, we resample all fields to a \( 1^\circ \times 1^\circ \) resolution and apply a common landmask to all products. In computing the mean field, we choose to include all monthly fields and treat nonexistent data as zero fluxes. For example, if a grid box is covered by sea ice for 10 out of 12 months but has a net heat input of 20 W m\(^{-2}\) during the remaining 2 months, we take the time-mean input to be \((20 + 20)/12 \approx 3.3\) W m\(^{-2}\). The riverine runoff compiled from Dai and Trenberth (2002) and Dai et al. (2009) is resampled to the closest point defined as ocean by the landmask, where it is added to the balance of \( E-P-R \).

It should be noted that the estimates of both \( Q_{\text{net}} \) and \( E-P-R \) are questionable in polar regions as well as in regions of sparse observations, such as the Southern Ocean. While we use global estimates for all calculations as far as it is provided by each product, we do not wish to imply that these estimates are well constrained in high latitudes.

### 3. Results

With these estimates of \( Q_{\text{net}} \) and \( E-P-R \), it is possible to evaluate the global area integral of density forcing to estimate the effects of the seasonal cycle and to estimate which areas of the ocean have the highest contraction potential due to cabbeling effects.

#### a. Global cabbeling rates

When Eq. (4) is integrated over the global ocean, a nonzero result emerges as noted by McDougall and Garrett (1992):

\[
\iiint_A F_a \, dA \neq 0. \tag{7}
\]

Upon evaluation of Eq. (7), using the balanced OAFlux dataset, we find a global imbalance of \( 7.6 \times 10^7 \) kg s\(^{-1}\), which translates to an equivalent volume flux of \( 74,000 \) m\(^3\) s\(^{-1}\) or 0.07 Sv, which is equivalent to a global sea level rise of 6.3 mm yr\(^{-1}\) or approximately 450% of the average discharge of the Mississippi River. This compares to observations between 1972 and 2008, in which a sea level rise of approximately \( 2.1 \pm 1 \) mm yr\(^{-1}\) was observed from tide gauges and satellite altimetry (Church et al. 2011). The spatial distribution of buoyancy flux (i.e., \( F_p \)) is shown in Fig. 3 and the zonal averages of the thermal component of the buoyancy flux (i.e., \( F_{pT} \)) for each degree of latitude are shown in Fig. 2.

All western boundary currents show a clear positive density flux, caused by the cooling (and hence densification) of relatively warm water that is advected poleward. This is also evident in the zonal averages of the heat flux into the ocean as shown in Fig. 2.

To illustrate the relationship between the surface forcing and the dissipation in the interior, we chose to focus on the leading terms and consequently ignore the effects of salinity and \( \beta \). As Joyce (1980) has shown, the interior dissipation of thermal variance is equal to the surface integral of the product of heat flux and temperature. That is, the heating of warm water and the cooling of cold water generates large-scale thermal variance that must be dissipated by interior mixing. Following Joyce, we obtain an expression relating the surface production and interior dissipation of thermal variance in the ocean:

\[
\frac{1}{\rho_0 c_p} \iiint_A (\bar{\theta} Q_{\text{net}}) \, dA = -\iiint_V \nabla \bar{\theta} \cdot \mathbf{u'} \, dV = \frac{1}{2} \iiint_V \chi_T \, dV, \tag{8}
\]

where overbars in Eq. (8) denote temporal averages and primes denote fluctuating components, such that \( \theta' = \theta - \bar{\theta} \) and \( \mathbf{u'} = \mathbf{u} - \bar{\mathbf{u}} \). Of course, to first order this result could be derived from Eq. (3) because the variations in \( \theta \) and \( \alpha \) are directly related (Figs. 1 and 2). When this expression is evaluated using a scale height of \( H = 600 \) m, Joyce (1980) found values of \( \chi_T = 1 \times 10^{-7} \) C\(^2\) s\(^{-1}\), while we estimate \( \chi_T \) as \( 8 \times 10^{-8} \) C\(^2\) s\(^{-1}\) using \( 1^\circ\)-resolution fields of all variables discussed in section 2 and also assuming a scale height of \( H_S = 600 \) m. This means that we assume most of the dissipation occurs in the upper 600 m of the ocean.

We can use these separate estimates of buoyancy production and dissipation rates as rough checks on one another. That is, we find that a volume-weighted average value of \( \partial a/\partial \theta \big|_{S_p} \) of \( 9.8 \times 10^{-6} \) C\(^{-2}\), as computed with the Gibbs Seawater (GSW) Oceanographic Toolbox (McDougall and Barker 2011), which uses the Thermodynamic Equation Of Seawater-2010 (TEOS-10) equation of state (Intergovernmental Oceanographic Commission 2011). This would correspond to a volume-weighted
potential temperature of 3.15°C. Then we compare this with the buoyancy input (in a steady-state ocean) of \( V \cdot \mathbf{n} = 5.6 \times 10^{-14} \text{ s}^{-1} \) or \( 7.6 \times 10^{-14} \text{ s}^{-1} \) when seasonal effects are included, equivalent to a theoretical sea level rise of 6.3 mm yr\(^{-1}\) or 8.6 mm yr\(^{-1}\), respectively.

When scaling \( \chi_T \) to the mean depth of the ocean instead of a scale height of \( H_S = 600 \text{ m} \), we obtain a value of \( \chi_{T\text{full}} = 1.3 \times 10^{-8} \text{ °C}^2 \text{ s}^{-1} \). When these values are substituted into the rhs of Eq. (1), we find \( V \cdot \mathbf{n} \) to be approximately \(-6.4 \times 10^{-14} \text{ s}^{-1}\). This value would be equivalent to a sea level change of \(-7.2 \text{ mm yr}^{-1}\), comparable (and, as expected, opposite in sign) to the value obtained from the buoyancy input rates. As this calculation assumes a spatially uniform dissipation of thermal variance, which may cause a high bias in this result because \( \frac{\partial \alpha}{\partial \theta} \big|_{\delta_p} \) would be lower in the upper ocean where most of the dissipation occurs, we repeat this calculation, assuming that the dissipation rate is proportional to the temperature gradient \( \delta_T \) or the square of the temperature gradient \( (\delta_T)^2 \). This results in \( \frac{\partial \alpha}{\partial \theta} \big|_{\delta_p} = 9.4 \times 10^{-6} \text{ °C}^{-2} \) and \( 8.6 \times 10^{-6} \text{ °C}^{-2} \), changing \( V \cdot \mathbf{n} \) to \(-6.1 \times 10^{-14} \text{ s}^{-1}\) and \(-5.6 \times 10^{-14} \text{ s}^{-1}\), respectively. Given the uncertainties in the datasets, these values closely match the inferred cabling from the buoyancy flux estimates.

Of course, the actual distribution of \( \chi_T \) as a function of temperature would have to be estimated to determine the distribution of cabling-induced contraction in the ocean. In section 3d, we make an estimate of where simple vertical mixing would cause the greatest contraction on mixing.

b. Effects of the seasonal cycle

So far, all calculations were performed using time means of all parameters. There is, however, another effect that was pointed out by Zahariev and Garrett (1997): because the peak of the heat flux roughly coincides with the seasonal temperature peak (and vice versa), a positive buoyancy flux is introduced. The inclusion of the seasonal cycle increases the global surface buoyancy forcing by 35% relative to the time-mean case to 101 000 m\(^2\) s\(^{-1}\), which is equivalent to a sea level rise of 8.6 mm yr\(^{-1}\) or 600% of the Mississippi River mean discharge. This is illustrated in Fig. 4 in which the zonal average of the thermal component of the buoyancy flux is shown using the time mean and using monthly data to include the effects of the seasonal cycle. There is little difference in the tropics because seasonality is suppressed and only small differences in midlatitude. However, the seasonal contribution is significant to the global sum.

It is likely that only a fraction of the seasonal input is mixed below the seasonal thermocline and dissipated there, with most of thermal variance being destroyed by seasonal mixing in the upper ocean through entrainment of deeper water. This was shown by Zahariev and Garrett (1997). Similarly, there are also nonzero effects due to the diurnal cycle, which causes the top few meters of the ocean to heat and cool. However, this is of a much smaller magnitude because diurnal variations of SST are considerably less than seasonal ones. This is also likely dissipated within the depth of the diurnal thermocline by entrainment of deeper water and is unlikely to have any significant effect on the ocean interior.

c. Intercomparison between datasets

There are tens of global evaporation, precipitation, and heat flux estimates available today. These products may be broadly classified as reanalysis products or observational datasets. While reanalysis products should be more self-consistent (which is not always the case, as shown for example by Sterl (2004)), comparisons with in situ buoy observations show broader agreement with observational datasets (e.g., Yu et al. 2008). Because a formal error analysis of a combination of such products becomes virtually impossible owing to unknown covariance and biases between the products, we instead choose to apply the same methodology that was used to generate density flux estimates with OAFlux and the Schanze et al. (2010) dataset with some of the most commonly used \( E, P, \) and \( Q_{\text{net}} \) products.

The products used are the Modern-Era Retrospective Analysis for Research and Applications (MERRA) (Bosilovich et al. 2006) (\( E, P, \) and \( Q_{\text{net}} \)), the adjusted Common Ocean–Ice Reference Experiment, version 2
(CORE.2), reanalysis (Large and Yeager 2009; NCAR 2010) (E and P), the European Centre for Medium-Range Weather Forecasts Interim Re-Analysis (ERA-Interim) (E, P, and \( Q_{\text{net}} \)), the National Oceanography Centre Southampton Flux Dataset, version 2.0 (NOCS v2) (Berry and Kent 2009) (\( Q_{\text{net}} \)), the Estimating the Circulation and Climate of the Ocean, Phase II (ECCO2) ocean state estimate (Menemenlis et al. 2008) (\( E-P-R \) and \( Q_{\text{net}} \)), the Japanese ocean flux datasets with use of remote sensing observations (J-OFURO) (Kubota et al. 2002) (\( Q_{\text{net}} \)) as well as two experimental versions of OAFlux with different shortwave and longwave (\( Q_{\text{SW}} \) and \( Q_{\text{LW}} \) products [Clouds and the Earth’s Radiant Energy System (CERES) and National Aeronautics and Space Administration (NASA)/Global Energy and Water Cycle Experiment (GEWEX) Surface Radiation Budget (SRB) Release-3.0 instead of the International Satellite Cloud Climatology Project (ISCCP), which is used in the currently published version] (L. L. Yu 2012, personal communication). The evaluated time period of the datasets is 1987–2004 except for ECCO2, which starts in 1990, and J-OFURO2, which starts in 1988.

The results of evaluating the time mean of all products to produce buoyancy flux estimates are shown in Table 1, while the buoyancy flux estimates including seasonal effects are tabulated in Table 2. While most datasets agree with each other to within approximately 50%, the MERRA \( Q_{\text{net}} \) product suggests more than double the buoyancy flux than the mean of the products; this is likely due to extreme heat fluxes at high latitudes. The lowest estimates are produced by the ERA-Interim and ECCO2 \( Q_{\text{net}} \) estimates. As may be expected from theory, the choice of the \( E-P-R \) dataset influences most results by less than 5% in the time-mean case and approximately 10% when the seasonal cycle is included.

d. Cabling due to vertical mixing

This estimate of the cabling rate poses the further question: if the ocean were entirely mixed through stirring, causing uniform conservative temperature \( \Theta \) and salinity, what would the resulting drop in sea level be?

This question was addressed by Gille (2004), who found a potential sea level height decrease of 165 mm upon complete (stirring) homogenization of the upper 3000 m of the ocean. We repeat the Gille experiment with the EN3 dataset to the full available depth to illustrate the potential contraction of the water column due to vertical mixing at each point in the ocean. For the purpose of computing density (and hence density changes) at each time step, we use the TEOS-10 equation of state (Intergovernmental Oceanographic Commission 2011) as implemented in the GSW toolbox (McDougall and Barker 2011) and use conservative temperature as the tracer instead of potential temperature to avoid issues of enthalpy that were pointed out by McDougall (2003). We thus allow the vertical profile at each 1° latitude and longitude to diffuse according to Eq. (9) with a mixing coefficient of \( \kappa_z = 1 \times 10^{-3} \text{m}^2 \text{s}^{-1} \) over a period of 10 000 years, motivated by Davis (1994) and Gille (2004), which is virtually an end state of turbulent mixing:

\[
\frac{\partial \Theta}{\partial t} = \kappa_z \frac{\partial^2 \Theta}{\partial z^2}. \tag{9}
\]

This equation is integrated numerically and the result is shown in Fig. 5. The greatest cabling potential is found in the subtropical gyres, particularly in the Pacific and Indian Oceans but also in the North Atlantic. From this it is evident that the contraction potential is predominantly a function of the main thermocline temperature gradient and depth as well as the total ocean depth at each grid point. By calculating the area-weighted average of the vertical cabling contraction potential, we estimate the global sea level drop that would result from vertical mixing as 230 mm.

e. Total cabling potential in the ocean

While we find that the global mean sea level would decrease by 230 mm solely by vertical mixing as outlined...

| \( Q_{\text{net}} \) | \( E-P-R \) | \( E-P-R \) |
|----------------|----------|----------|
| \( \text{OAJFlux (ISCCP)} \) | +6.3 | +6.3 | +6.4 | +6.3 |
| \( \text{OAJFlux (CERES)} \) | +5.9 | +5.8 | +6.2 | +5.9 |
| \( \text{OAJFlux (SRB)} \) | +8.8 | +8.7 | +9.0 | +8.8 |
| \( \text{ERA-Interim} \) | +3.4 | +3.3 | +3.9 | +3.3 |
| \( \text{MERRA} \) | +11.7 | +11.6 | +12.3 | +11.6 |
| \( \text{ECCO2} \) | +4.3 | +4.2 | +4.5 | +4.3 |
| \( \text{NOCS v.2} \) | +6.4 | +6.4 | +7.3 | +6.4 |
| \( \text{J-OFURO2} \) | +9.6 | +9.5 | +10.3 | +9.5 |

| \( Q_{\text{net}} \) | \( E-P-R \) | \( E-P-R \) |
|----------------|----------|----------|
| \( \text{OAJFlux (ISCCP)} \) | +8.6 | +9.6 | +9.2 | +8.8 |
| \( \text{OAJFlux (CERES)} \) | +7.9 | +8.9 | +8.7 | +8.1 |
| \( \text{OAJFlux (SRB)} \) | +11.0 | +12.0 | +11.8 | +11.2 |
| \( \text{ERA-Interim} \) | +5.2 | +6.2 | +6.3 | +5.4 |
| \( \text{MERRA} \) | +13.5 | +14.4 | +14.7 | +13.7 |
| \( \text{ECCO2} \) | +5.7 | +7.1 | +6.6 | +6.0 |
| \( \text{NOCS v.2} \) | +8.4 | +9.4 | +9.8 | +8.6 |
| \( \text{J-OFURO2} \) | +12.0 | +13.0 | +13.3 | +12.2 |
in the previous section, the elimination of horizontal temperature and salinity gradients would result in even more cabbeling. We thus homogenize all available cells of the EN3 dataset to an end state caused by turbulent mixing using the GSW toolbox (McDougall and Barker 2011) as before. Turbulent mixing would lead to a homogenization of conservative temperature rather than in situ temperature, and thus the end state is one of uniform salinity and conservative temperature. While one may speculate about a molecular-diffusive end state, this is not considered here, as it would take millions of years of no turbulent mixing to achieve this, making this a purely fictional exercise. Further, it should be noted that the final equilibrium state of a quiet ocean would actually have a strong increase in salinity with depth as it reached diffusive balance with the gravitational field due to the baro-diffusion effect (Levenspiel and de Nevers 1974; Intergovernmental Oceanographic Commission 2011). This would only happen on geological time scales (and in the absence of turbulent mixing) and is not considered as a plausible end state.

All computations are performed using the TEOS-10 equations to full precision using the GSW toolbox (McDougall and Barker 2011). The total cabbeling potential after full horizontal and vertical homogenization to uniform salinity and conservative temperature would cause the sea level to decrease by 300 mm. This shows that, while vertical mixing in the ocean is the dominant contributor, a significant fraction, $\sim 1/4$, of the global ocean cabbeling potential can be attributed to horizontal mixing.

4. Summary

New estimates of surface buoyancy fluxes due to heat and moisture transport have allowed us to estimate the global integral of density forcing. Because the ocean is heated predominantly in warm areas with a higher thermal expansion coefficient and cooled in cold areas with a low expansion coefficient, an addition of buoyancy (a net density loss due to expansion) is estimated. As McDougall and Garrett (1992) and Davis (1994) have suggested, this loss of density must be compensated by the interior densification on mixing due to the nonlinear equation of state (cabbeling). We show that the estimated cabbeling is consistent with expected dissipation rates for thermal variance as in Joyce (1980). We find that the mean density flux would contribute to a sea level rise of 6.3 mm yr$^{-1}$ (0.07 Sv) if uncompensated by mixing. The correspondence between surface temperature and heat flux in the seasonal cycle provides an additional 35% of potential sea level rise (increasing it to 8.6 mm yr$^{-1}$ or 0.1 Sv). These numbers correspond to volume fluxes that are 4.5 and 6 times larger than the average discharge of the Mississippi River, respectively.

While the choice of the data product does influence this result to within a factor of $\sim 2$, it is clear that significant mixing and, therefore, cabbeling is occurring in the ocean interior to absorb this apparent expansion of the ocean. We find that an independent estimate of the velocity divergence (i.e., $\nabla \cdot \mathbf{v}$) based on the dissipation of thermal variance provides a close match to our estimate of the time-mean cabbeling rate as inferred from the surface density forcing. The potential sea level change upon complete turbulent vertical mixing of the ocean is estimated as 230 mm. When horizontal mixing is included, the global ocean cabbeling potential is estimated for a turbulent mixing end state as 300 mm. While this number will never occur in today’s climate system, it illustrates that changes in the stratification or mixing rate of the ocean may play a nontrivial role in global sea level change.

Also, the question arises of whether variations in ocean mixing rates can contribute to the temporal variability observed in sea level records (Church et al. 2011). An evaluation of the interannual variability of the surface density flux $F_{\rho}$ from $Q_{\text{net}}$ and $E-P-R$ products discussed here suggests variabilities ranging from approximately 1 to 3 mm yr$^{-1}$. While there are large uncertainties associated with these estimates, particularly in $Q_{\text{net}}$ products, this signal is potentially relevant to studies of sea level variability, especially if significant variations in mixing rates occur due to (say) year-to-year changes in storminess, the El Niño–Southern Oscillation or the 18.6-yr cycle in tides (e.g., Loder and Garrett 1978).

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