Vertical mixing and coherent anticyclones in the ocean: The role of stratification

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Abstract. The role played by wind-forced anticyclones in the vertical transport and mixing at the ocean mesoscale is investigated with a primitive-equation numerical model in an idealized configuration. The focus of this work is to determine how the ocean stratification impacts such transport.

The flows, forced only at the surface by an idealized wind forcing, are predominantly horizontal and, on average, quasi-geostrophic. In the vortex cores and inside intense filaments, however, the dynamics is strongly ageostrophic.

Mesoscale anticyclones appear as ”islands” of increased penetration of wind energy into the ocean interior, and they represent the maxima of available potential energy. The amount of available potential energy is directly correlated with the degree of stratification.

The wind energy injected at the surface is transferred at depth through the generation and subsequent straining effect of Vortex Rossby Waves (VRWs), and through near-inertial internal oscillations trapped inside anticyclonic vortices. Both these mechanisms are affected by stratification. Stronger transfer but larger confinement close to the surface is found when the stratification is stronger. For weaker stratification, vertical mixing close to the surface is less intense but below about 150 m attains substantially higher values due to an increased contribution of both VRWs, whose time-scale is on the order of few days, and of near-inertial motions, on time scale of few hours.

1 Introduction

The ocean circulation is characterized by a surface wind driven component, a stable stratification below the first few hundred meters of the water column, a meridional overturning circulation and turbulent motions. Turbulence in the ocean encompasses many dynamical scales, but at the mesoscale, between 10 and 500km, the motion is quasi two-dimensional and characterized by the presence of coherent vortices. They permeate many oceanic regions and form primarily by internal instability of the flow field along the baroclinically-unstable meanders of boundary currents and jets, locally through the interaction of the flow with the topography, or as a response to the atmospheric forcing that acts as eddy kinetic energy source in oceanic regions far from strong currents (Stammer, 1997; Bracco and Pedlosky, 2003; Spall et al., 2007; Wunsch and Ferrari, 2004).

Eddies are crucial to the local dynamics and have important consequences on tracer dispersion and ocean stirring and mixing processes (Pasquero et al., 2007). Their role on the horizontal motion has been studied extensively in the framework of 2d-barotropic, quasigeostrophic, and shallow water turbulence (McWilliams and Weiss, 1994; Provenzale, 1999; Bracco et al., 2004; Pasquero et al., 2004, e.g.). On the other hand, the vertical structure and associated mixing of mesoscale vortices had long been viewed and parameterized in a simplified manner (Gent and McWilliams, 1995; Griffies, 2000). Recent findings, though, have highlighted their role in transferring wind-forced near-inertial oscillations into the ocean interior (Kunze, 1985; Klein et al., 2004; Darniaux et al., 2008). In this respect they constitute a different pathway by which inertial energy can reach the ocean interior and their contribution to vertical mixing may be fundamental in maintaining the meridional overturning circulation. In particular, cyclonic and anticyclonic eddies expel or trap, respectively, inertial oscillations as shown by Kunze (1985) for quasigeostrophic flows. Following these works, in Koszalka et al. (2009) it has been showed that in a strongly stratified ocean the vertical circulation associated with wind-forced anticyclones is one order of magnitude more intense than predicted by the quasigeostrophic omega equation and exhibits a complex spatial pattern, akin to one reported for frontal regions. At local scales, the strong and complex ver-
tactical mixing associated with the eddies may help interpreting recent observations of biological fields (Benitez-Nelson et al., 2007; McGillicuddy et al., 2007).

In this work we use the idealized set-up of Koszalka et al. (2009) and high vertical resolution to investigate with a primitive equation model how stratification impacts the representation of vertical transport and mixing with a wind-driven mesoscale eddy field. The horizontal and vertical extension of oceanic eddies is indeed affected by the local stratification (Smith and Vallis, 2001, 2002; Arbic et al., 2007). However, while it has been shown that the ambient stratification does not to affect the horizontal transport and mixing properties of mesoscale vortices (Bracco et al., 2004), its influence in setting the vertical ones has not yet been investigated in details.

In light of the recent findings on wind-forced near-inertial oscillations, we have chosen a flow configuration that insures a predominance of anticlones. Dominance of anticlones has been found in a number of numerical and theoretical solutions of the geostrophic turbulence that are representative of the ocean interior (Polvani et al., 1994; Arai and Yamagata, 1994; Cho and Polvani, 1996; Yavneh et al., 1997). Here we consider two different vertical stratification profiles and we analyze the eddy population that emerges in an idealized, wind-forced portion of the open ocean. We study the transfer of the wind energy into the underlying flow and its redistribution in the fluid column, focusing on the vertical transport properties of the mesoscale vortices.

2 Numerical model

In this study we use the Regional Ocean Modelling System (ROMS). ROMS is an incompressible, free-surface, hydrostatic, primitive-equations, circulation model (Shchepetkin and McWilliams, 2005) with a generalized vertical, terrain-following, coordinate system (s-coordinate). The idealized configuration implemented is similar to the one described in detail in Koszalka et al. (2009). The domain is doubly periodic, with a constant depth $H = 1000$ m and lateral size $L = 256$ km. The resolution is $\Delta x = 1$ km in the horizontal and 80 layers in the vertical, 23 of which confined in the upper 100 meters of the water column, respectively. The Coriolis frequency is $f = 10^{-4}$ s$^{-1}$. In the ocean, the degree of stratification is described by the buoyancy frequency, $N(z) = -\frac{\partial \rho}{\partial z}$, where $z$ is a local vertical coordinate and $\rho$ is potential density. In most oceanic regions $N/f$, which measures the relative importance of stratification and rotation, is greater than 1, and is typically of order 10 or more, with larger values attained near the surface than at depth.

Here we consider two vertical stratifications, as shown in Fig. 1. In the following we refer to the strongest stratified case as SS and to the weaker one as WS. The SS profile is similar to the one in Koszalka et al. (2009) but for the absence of the mixed layer. The first internal Rossby radius of deformation, given by $L_I = \frac{\sqrt{\pi N H}}{f}$ (Vallis, 2006) is $L_I^{SS} \approx 15$ km for SS and $L_I^{WS} \approx 11$ km for IS. The (barotropic) Rossby radius of deformation is $L_o = \sqrt{gH/f} \approx 1000$ km in both configurations.

Biharmonic horizontal diffusion acts with coefficient $A_H = 10^6$ m$^4$s$^{-1}$ on all fields and runs, and the non-dimensional quadratic bottom drag parameter is $3 \times 10^{-4}$. A non-local, K-Profile Parameterization (KPP) scheme (Large et al., 1994) is used to parameterize vertical mixing. The scheme modifies the background value of vertical mixing coefficient, $K_{v,coef} = 10^{-6}$ m$^2$s$^{-1}$ in both runs, in the surface boundary layer by employing the Monin-Obukhov similarity theory, and in the interior by accounting for contributions due to the shear instability as function of the local Richardson number, and due internal wave breaking assumed to be inversely proportional to $N$. The vertical dissipation is thus function of the system state and will vary between the configurations.

The fluid is initially at rest and it is constantly forced by a spatially-variable, narrow-band sinusoidal wind stress of amplitude 0.1 N m$^{-2}$, centered on the wavenumber $k_x = k_y = 3$, i.e. with a radial wavenumber $k_f = \sqrt{k_x^2 + k_y^2} \approx 4$ and length scale $L_f \approx 64$ km. This idealized forcing is used as an artifice to maintain the mesoscale turbulence in the homogeneous domain in lieu of the inclusion of inhomogeneous mean-flow instability processes. It was verified in Koszalka et al. (2009) that the evolution and transport properties of the simulated flow do not depend critically on the detail of the forcing field for a given radial wavenumber. The wind forcing is identical in all configurations. Radiative forcing is not included in these idealized runs.

3 Results

3.1 Wind-forced vortices

Within the time scale of the inertial period, the wind forcing sets up a circulation in the Ekman layer with depth approximately given by $d_E = \frac{2\Delta K u}{\Delta f}$ (Vallis, 2006), where $<Ku>$ is the domain-averaged vertical mixing coefficient for momentum in the KPP scheme. In all runs $d_E$ is about 6 m.

After about 200 days, the system reaches a statistical stationary state, on which we focus our analysis. The corresponding domain-averaged profiles of temperature, salinity, density, Brunt-Vaisala frequency and kinetic energy are shown in Fig. 1. The dynamics is surface-intensified in both cases due to the chosen forcing. The stratification influences the depth and strength of the energy transfers and this results in the near-surface flow in strongly-stratified case SS.

\footnote{We follow Vallis (2006) and use the definition without inclusion of the factor of $\pi$. The estimates given by the normal mode analysis (dynmodes.m of J. Klinck) give slightly higher values, $L_I^{SS} \approx 20$ km and $L_I^{WS} \approx 15$ km.}
being twice that strong than in WS and slightly weaker at mid-depth (ca. between 220-350m).

The flow is dominated by the presence of coherent vortices, predominantly anticyclonic, as shown by the vertical component of the relative vorticity, \( \zeta = \partial_x v - \partial_y u \), in Fig. 2a-b. The vortices are long-lived (up to several months) and undergo complex evolution involving perturbations under the straining field due to wind forcing, instabilities in the vortex cores and merging events.

The asymmetry in the vortex population in stratified flows with small \( L_I \) and finite \( Ro \) has been discussed in Koszalka et al. (2009) and is linked to the straining field exerted by vortex Rossby waves (VRWs), which are evident in the presence of strong cyclonic filaments close to the surface. The depth of maximum asymmetry is linked to the penetration depth of the vortex cores. Below the vortices the two cases namely, the higher amount of surface baroclinic energy found in the SS case (Fig. 3c) appears to be concentrated at scales larger than the forcing; the WS configuration is indeed characterized by smaller \( L_I \) and thus stronger inhibition of the transfers in the inverse energy cascade range.

Recently Arbic et al. (2007) considered a two-layer quasi-geostrophic model to investigate the dynamics of Ekman-damped, turbulent oceanic flows with surface-intensified stratification, and found that the baroclinic energy is spatially localized at horizontal scales near \( L_I \) while the barotropic energy dominates the spatial average, in qualitatively agreement with the results presented here.

To quantify how mesoscale vortices and submesoscale filaments affect vertical transport in the ocean we first analyze the conversion of energy injected at the surface by the idealized wind. The transfer of wind energy into the ocean can be quantified by the wind power, i.e. the mean rate of the wind work, given by the mean product of the wind stress and the surface velocity, \( WW = \int \int \tau u \mathrm{d}x \mathrm{d}y \) (e.g. Wunsch and Ferrari, 2004; Brown and Fedorov, 2008). In both runs, forced by identical wind forcing, the mean wind power per unit area is positive and approximately \( WW = 0.001W/m^2 \). The standard deviation of the two time-series, however, differs quite significantly (7 \( \cdot \) \( 10^{-3} \) for SS and 4 \( \cdot \) \( 10^{-3} \) for WS) and so does the low frequency variability, with periodicity of about 30 days for SS and 15-20 days for WS (Fig. 5a).

Only a portion of the energy introduced by the wind is dynamically active and can be transformed in kinetic via baroclinic instability. This is the available potential energy of the system, defined as \( APE = \int \int \rho^2 / \gamma^2 / 2 > dxdydz \), where \( \rho \) is the deviation from the mean density profile and \( \gamma^2 = N^2 / (g^2 / \rho_o) \) (Oort et al., 1989). The APE differs significantly between the two runs, and is twice as strong in the SS case (Fig. 5b). The vortices correspond to local maxima in APE (Fig. 5c). They cover only 5\% of the domain area in both SS and WS runs, but they account for about 65\% of the total APE in both cases. Thus, given the different levels of APE, the role of vortices in the energy conversion
increases with increasing strength of the stratification.\(^2\).

The presence of vortices and filaments may further affect the vertical transport in two ways: By modifying the vertical mixing coefficient used in the KPP parameterization scheme, and by inducing a vertical circulation. Those two mechanisms are separately investigated below.

### 3.2 Vertical mixing due to the presence of the vortices

The vertical profiles of the surface average effective vertical mixing term \(Kv\) determined by the KPP scheme (Large et al., 1994) are shown in Fig. 6a. The mean value of \(Kv\) does not vary significantly with stratification, being only modestly weaker between 100 and 200 m in the SS case, and features a common vertical structure. The constant background value of \(Kv_{cyclic}\) is magnified in the surface boundary layer. Fig. 6b shows a horizontal slice of \(Kv\) at 30m for the SS run (analogous results are found in the WS case). Near the surface, due to locally strong vertical shear and reduced stratification, the anticyclonic vortices act like “islands” of enhanced vertical mixing, and thus enhanced transfer of momentum into deeper layers. \(Kv\), which is inversely proportional to the ambient value of \(N\), increases again below 300m, where \(N\) is small (Fig. 1) and conditions are favorable to internal-wave breaking phenomena that the scheme aims at parameterizing (Large et al., 1994).

### 3.3 Vertical circulation

A snapshot of the instantaneous vertical velocity field at 80m is shown in fig.2c-d for the SS and WS cases, respectively. Vertical velocities are the most intense in the vicinity of the vortices and reveal a complex small scale structure. The vertical velocity variance profiles (averaged over 100 days, Fig. 7a) reveal that the maximum of \(w\) increases with increasing magnitude of the stratification; The SS case exhibits the strongest vertical velocities close to the surface with maximum values attained at about 60m. This is related to an enhancement of the momentum transfer within the vortex cores that are constrained in the first few hundred meters, as inferred from horizontal kinetic energy spectra. Few vortices can seldom penetrate deeper and induce high vertical velocities down to 600m. The weaker stratification case WS displays smaller vertical velocities close to the surface and a shift of the maximum to deeper layers (160m), due to the increased energy transfer in the vertical. Below this maximum, however, the WS run displays higher velocities than SS through the remaining of the water column, with a core of high values localized between 100 and 500m of depth. The overall structure of the velocity field variance is quite different in the two cases, despite a relatively small change in stratification.

The overall vertical structure of the vertical velocity field (including the exact positions of the depth maxima) features a low-frequency variability strongly linked to the dynamics of the anticyclonic vortices dominating the flow (Fig. 7b-c), as indicated by the Hovmöller diagrams of velocity variance and relative vorticity anomalies. This suggests an important role for intrinsic ocean dynamics at sub-monthly scales in regions of high eddy activity.

The wavenumber spectra of the vertical velocity (Fig. 8a) confirms that close to the surface vertical velocities are higher in the SS case, at scales both larger and smaller than the forcing. Below 160m the WS flow displays larger vertical velocities than the SS one, but contributions are only from scales at or below the forcing.

The frequency spectra of \(w\) (Fig. 8b) reveal the presence of near-inertial internal waves in the deeper layers, broadly in correspondence to the position of the vertical velocity maximum in Fig. 7a). The near-inertial peak is shifted towards higher frequencies. This results from refraction by the eddy relative vorticity and trapping of waves within anticyclonic vortices, and is associated with the reduction of the horizontal scales of the waves and subsequent transfer of the wave momentum to deeper layers (Kunze, 1985; Danioux et al., 2008).

This trapping and transfer of wave momentum lead to the emergence of a maximum in \(\sigma_w\) at depth in agreement with the analytical and numerical investigation by Danioux et al. (2008). In their study, focused on an unstable jet forced by time-varying but spatially-uniform winds, the deep maximum was captured by the lowest vertical normal modes. They find that the signature of near inertial motions related to the first mode is a peak in the frequency spectra close to twice the inertial frequency \(2f_p\) corresponding to energy concentrated near the wavenumber \(\sqrt{3/L_I}\). This is consistent with our results for both SS and WS; the peaks in our work become broader due to the dispersion of the waves. The emergence of this scale has been explained analytically by using a shallow-water model in Danioux and Klein (2008) as a local resonance in the wavenumber space, following the excitation by the eddy relative vorticity and involving the non-linear term \(v\partial_yh\), where \(h\) is the free-surface. In our fully three-dimensional primitive-equation model, this translates into \((u\partial_yw, v\partial_yw)\) terms in the momentum equation and corresponds to the “tilting term” in eq. 1 below.

Near the surface the vertical velocity is dominated by low(er)-frequency motions. Vortices at different scale set the spectral slope and the overall behavior of the vertical velocities. Their contribution is modulated by the presence of VRWs that imprint their signature with a peak more evident in the WS case and centered near \(1/(fp/5.8) \approx 4\) day, corresponding to the period of these waves given the \(L_I\) con-

\(^2\)To quantify the area covered by the vortices we used the Okubo-Weiss parameter \(OW = S^2 - \zeta^2\), where \(S^2\) is squared strain, and \(\zeta\) is vorticity (Weiss, 1981). \(OW < -\sigma_{ow}\) identifies vortex cores and \(OW > \sigma_{ow}\) characterizes the cyclonic rings around the vortices and the strongest filaments. For a discussion on the use of \(OW\) in OG and PE flows see (Petersen et al., 2006; Koszalka et al., 2009)
sidered (Koszalka et al., 2009). The higher contribution of the frequencies in the VRW-range in the WS case is consistent with the \( L_I \) being smaller and is a consequence of the importance of the straining effect of VRWs on the overall dynamics, that increases for smaller \( L_I \) (Graves et al., 2006; Koszalka et al., 2009).

At 200 m and 600 m depth both contributions of near-inertial motions and VRWs are significantly more important in the WS case, suggesting that mixing below the surface is strongly enhanced for reduced stratification through those two mechanisms. Near-inertial waves trapped within anticyclones remain stronger in the WS case at all depths (see Fig. 2e-f).

The physical processes contributing to the vertical velocity field, VRWs close to the surface, and near-inertial waves below the first 100 m, can be identified also considering the \( w \) diagnostic equation derived in Koszalka et al. (2009) and here repeated (see their Appendix for full derivation):

\[
\begin{align*}
\text{FREE SURFACE:} & \quad w(x, y, z) = \frac{D\eta}{Dt} \\
\text{ADVECTIVE:} & \quad - \int_z^\eta \alpha_1 \left[ \frac{\partial \xi_1}{\partial t} + u \frac{\partial \xi_1}{\partial x} + v \frac{\partial \xi_1}{\partial y} + w \frac{\partial \xi_1}{\partial z} \right] d\eta \\
\text{STRETCHING:} & \quad - \int_z^\eta \alpha_1 \left[ \chi_1 \xi_1 \right] d\eta - \int_z^\eta \alpha_2 \left[ \chi_1 \xi_2 \right] d\eta \\
\text{TILTING:} & \quad - \int_z^\eta \alpha_1 \left[ \frac{\partial w}{\partial x} \frac{\partial \eta}{\partial x} + \frac{\partial w}{\partial y} \frac{\partial \eta}{\partial y} \right] d\eta + \int_z^\eta \alpha_2 \left[ \frac{\partial w}{\partial y} \frac{\partial \eta}{\partial y} \right] d\eta \\
\text{WIND STRESS:} & \quad + \int_z^\eta \frac{\alpha_2}{\rho_o} \left[ \frac{\partial}{\partial y} \frac{\partial \tau_x}{\partial y} \right] d\eta \\
\text{HORIZONTAL MIXING:} & \quad + \sum_{i=1,2} \int_z^\eta \alpha_i A_H \left( \frac{\partial^2 \xi_i}{\partial x^2} + \frac{\partial^2 \xi_i}{\partial y^2} \right) d\eta \\
\text{VERTICAL MIXING:} & \quad + \int_z^\eta \alpha_1 \left[ \frac{\partial}{\partial x} \frac{\partial}{\partial z} \left( K_v \frac{\partial \eta}{\partial z} \right) \right] d\eta - \int_z^\eta \alpha_2 \left[ \frac{\partial}{\partial y} \frac{\partial}{\partial z} \left( K_v \frac{\partial \eta}{\partial z} \right) \right] d\eta,
\end{align*}
\]

where \( \xi_1 = \frac{\partial v}{\partial x}, \xi_2 = -\frac{\partial u}{\partial y}, \chi_1 = \frac{\partial u}{\partial x}, \chi_2 = \frac{\partial v}{\partial y}, \alpha_1 = (f + \xi_1)^{-1} \) and \( \alpha_2 = (f + \xi_2)^{-1} \).

In both runs the nonlinear advective (or ageostrophic) and stretching terms dominate. These two terms appear in the form of azimuthal wavenumber \( k_A = 4 \) disturbances around the vortex cores, are strongly anticorrelated and largely cancel each other (see Koszalka et al. (2009) and their Fig. 4). As a result, the tilting term dominates the overall pattern of \( w \) near the vortex cores in the upper 150m. The amplitude of the vertical velocity field, however, is dominated by the regions where the nonlinear advective (or ageostrophic) and stretching terms do not cancel out. These three dominant contributions when averaged over the inertial period bear very clearly expression of the VRWs, characterized by a sub-inertial timescales of 3-4 days (Graves et al., 2006; Koszalka et al., 2009). In the upper 100 m the contribution of the vertical mixing term is also significant and reaches 10m/s with a spatial variability is clearly related to the presence of the vortices (Koszalka et al., 2009).

The vertical profiles of the dominant contributions are shown in Fig. 9a-d for the SS and WS cases in the upper 500 m of the water column. The vertical mixing term is comparable in the two runs, while the tilting term is larger below the first 100 m for weaker stratification, due to the higher internal wave activity in the WS case already seen in the frequency spectra. Advection and stretching contributions, on the other hand, are directly linked to the stratification and are consistently stronger (but also closer in amplitude and thus more prone to cancellation) in the SS case below 50 m.

4 Discussion and Conclusions

The maintenance of the ocean circulation requires energy input from the atmospheric wind field, and the wind forcing is believed to be the main driving force for the mesoscale eddies far from the ocean boundaries (Stammer, 1997). Winds acting on the sea surface produce direct conversion of atmospheric kinetic energy into oceanic kinetic and potential energies. A fraction of this input feeds into the large-scale general circulation, partly removed by baroclinic instability whereby the mesoscale flow structures arise. However the pathways of its further transmission to increasingly small scales in the ocean interior, and the role of mesoscale circulation in this transmission, are still uncertain (Wunsch and Ferrari, 2004; Brown and Fedorov, 2008). Inertial motions are the most obvious candidate for small-scale mixing in the ocean, with various possible generation mechanisms.

Recent works point to mesoscale vortices as responsible for transferring wind energy into the ocean interior through near inertial oscillations (Kunze, 1985; Klein et al., 2004; Danioux et al., 2008). Here, with a primitive-equation numerical model in an idealized set-up, we investigated the role of wind-forced anticyclones in the vertical mixing and the dependence of the (modeled) mesoscale ocean circulation on stratification, considering two different stratification profiles.

The flows are forced by an identical wind forcing and the resulting circulation is predominantly horizontal, bearing strong resemblance to that of quasigeostrophic turbulent flows. The circulation is indeed quasigeostrophic on average in both runs, featuring small \(< Ro >\), negligible contribution of divergence and predominance of horizontal kinetic energy component over the vertical one. In the vortex cores and inside intense filaments, however, dynamics is locally ageostrophic and the strength of the ageostrophic component increases with the strength of the stratification.
We showed that mesoscale anticyclones are "islands" of increased penetration of wind energy into the ocean interior, representing the maxima of available potential energy. The APE is, in turn, a function of stratification and increases with it.

The vertical transfer of wind energy happens due to an interplay between the local effect of the vortices that modify the density surfaces, the generation and straining effect of the Vortex Rossby Waves, and near-inertial internal oscillations trapped inside anticyclones (Kunze, 1985; Danioux et al., 2008; Koszalka et al., 2009).

In our runs these physical mechanisms appear to be significantly affected by stratification. We find that the vertical transfer of wind momentum - the "chimney effect of eddies", (Lee and Niiler, 1998) - is more pronounced close to the surface when the stratification is stronger and that penetration below the first hundred meters is limited to sporadic events. Differences linked to stratification are apparent when evaluating the vertical velocities and the associated mixing. For the SS case, the Vortex Rossby waves and their interactions with the Ekman circulation induce a more vigorous and spatially-complex $w$ field near the ocean surface that drives a vertical transfer of momentum and tracers, but VRWs are weaker and contain less energy below the first 150 m. For weaker stratification, vertical mixing close to the surface is less intense but below 150 m attains higher values due to a largely increased contribution of both VRWs, whose timescale is on the order of few days, and of near-inertial motions, on time scale of few hours.

In a turbulent flow dominated by the presence of coherent anticyclones, as it may be the ocean interior away from boundaries and strong currents, the overall diffusivity below the first 100 m tends therefore to increase with decreasing density difference. On time scales much longer than considered here, the intensified transfer of near-inertial momentum at depth could then enhance the mixing and potentially weaken further the stratification in a positive feedback loop. More work is needed to test this possibility in realistic configurations, but the implications for climate change scenarios could be severe.

This work was originally motivated by the idea that the ocean circulation in the Eocene could have been dominated by eddies in presence of a very weak thermocline Hay et al. (2004). We wanted to explore, in a idealized and simple configuration, how the vertical transport within eddies would be affected by stratification. We find that internal wave motion (VRWs and near-inertial contributions) could play an important role in the maintenance of a weak thermocline, through physical mechanisms quite different from what hypothesized by simple quasigeostrophic arguments.

As a note of caution, the reader should be reminded that in our model the vertical mixing coefficient, $K_v$, depends on forcing, stratification and velocity field and is calculated by the KPP scheme (Large et al., 1994). As a result of the interplay between the local stratification properties and the eddies, surface values of $K_v$ are very large in the vicinity of the anticyclonic vortices, enhancing their role in the vertical transfer of momentum and tracers into the deeper layers. The results presented may depend on the parameterization scheme. However, the KPP scheme is regarded as a reliable representation of the physics of the vertical mixing in ocean models and is widely implemented (e.g., Li et al., 2001).

In the view of the present results on the role of wind-forced mesoscale vortices in the transmission of the wind energy into the ocean and vertical transport, we raise a question whether the ocean circulation models, neither properly resolving mesoscale- and submesoscale dynamics, nor providing for spatial variability in the mixing, have any relevance to the study of the ocean circulation under changing climates.

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References


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Fig. 1. Domain-averaged profiles in the two configurations WS and SS, time-averaged over 100 days of (from left to right) temperature, salinity, density, Brunt-Vaisala frequency and kinetic energy.
Fig. 2. Instantaneous snapshots of relative vorticity at surface (10m depth) scaled by the Coriolis parameter, $\zeta/f$ for (a) SS and (b) WS case. Instantaneous vertical velocity field $w(\text{m day}^{-1})$ in (c) SS and (d) WS runs. Instantaneous vertical sections of $w(\text{m day}^{-1})$ through vortices in (e) SS and (f) WS, respectively.
Fig. 3. Surface-averaged vertical profiles: (a) skewness of the vertical component of the relative vorticity $\zeta$ (b) ratio of vertical component of the relative vorticity $\zeta$ to the horizontal divergence $\chi$, (c) the ratio of baroclinic (depth-dependent) component of the horizontal kinetic energy to the barotropic one. All statistical measures are averages over time period of 100 days.

Fig. 4. Top: Horizontal wavenumber spectra of the horizontal kinetic energy, time-averaged over 70 days: (a) barotropic (depth-averaged) component, (b) baroclinic (depth-dependent) component.
Fig. 5. Time series of the mean wind power $WW (W m^2)$ (a) and available potential energy $APE (J m^3)$ (b). A snapshot of $APE (J m^3)$ for SS run.
Fig. 6. (a) Surface-averaged profiles of the vertical mixing coefficient for momentum for the two simulations, time-averaged over 100 days. (b) A snapshot of vertical mixing coefficient at 30m depth for the SS case (units are $m^2/s^{-2}$.)
Fig. 7. (a) Surface-averaged vertical profiles of $\sigma_w$ for varying degree of stratification, time-averaged over 100 days. The position of the maxima are marked with dashed lines. (b)-(c) Hovmöller diagrams ($dt=1$ day) of surface-averaged profiles of the vertical velocity (top) and deviation from the mean r.m.s of relative vorticity (bottom) for SS and WS runs, respectively.
Fig. 8. (a) Horizontal wavenumber spectra of the vertical velocity in both experiments at 20m and 300m. Thick black lines mark wavenumber $\sqrt{3}/L_{II}^{SS}$ and black dashed lines correspond to wavenumbers $\sqrt{3}/L_{II}^{SI}$. (b) Frequency spectra of the vertical velocity at 20m, 200m and 600m.

Fig. 9. Variance of one-day averaged contributions in eq. 1 in both SS and WS cases corresponding to the fields shown in Fig. 2. (a) The vertical mixing contribution, (b) the tilting term, (c) the advective or ageostrophic component, and (d) the stretching term.