Energetic submesoscales maintain strong mixed layer stratification during an autumn storm

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ABSTRACT

Atmospheric storms are an important driver of changes in upper-ocean stratification and small-scale (1-100 m) turbulence. Yet, the modifying effects of submesoscale (0.1-10 km) motions in the ocean mixed layer on stratification and small-scale turbulence during a storm are not well understood. Here, we use large-eddy simulations to study the coupled response of submesocales and small-scale turbulence to the passage of an idealized autumn storm, with a wind stress representative of a storm observed in the North Atlantic above the Porcupine Abyssal Plain. Due to a relatively shallow mixed layer and a strong down-front wind, existing scaling theory predicts that submesoscales should be unable to re-stratify the mixed layer during the storm. In contrast, our simulations reveal a persistent and strong mean stratification in the mixed layer both during and after the storm. In addition, we find that the mean dissipation rate remains elevated throughout the mixed layer during the storm, despite the strong mean stratification. These results are attributed to strong spatial variability in stratification and small-scale turbulence at the submesoscale and have important implications for sampling and modeling submesoscales and their effects on stratification and turbulence in the upper-ocean.

1. Introduction

The upper ocean, particularly at mid-latitudes, is subject to intense, highly variable winds associated with synoptic atmospheric storms. These intermittent events energize nearly-isotropic turbulence at length scales smaller than the mixed layer depth, which drives entrainment and mixing of pycnocline water into the mixed layer and thereby deepens the mixed layer and increases its density (e.g. Davis et al. 1981; Large and Crawford 1995; Dohan and Davis 2011). In aggregate, storm-driven small-scale turbulence contributes significantly to the seasonal increase in the mixed layer depth and mixed layer density during the autumn in mid-latitudes (e.g. Large et al. 1986). Many previous studies have examined the upper-ocean response to storms using a one-dimensional framework (e.g. Pollard et al. 1972; Niiler and Kraus 1977; Price et al. 1978; Large et al. 1994). However, the upper ocean contains lateral variability associated with large-scale fronts, filaments, and eddies, which modify the evolution of upper-ocean stratification and small-scale turbulence during a storm.

Among the motions inducing lateral variability are submesoscales, anisotropic features with vertical scales similar to the mixed layer, horizontal scales between 0.1-10 km, and O(1) vorticity Rossby numbers (e.g. Thomas et al. 2008; Capet et al. 2008; McWilliams 2016), which are prevalent in the upper ocean (e.g. Munk et al. 2000; Shcherbina et al. 2013; Buckingham et al. 2016; Thompson et al. 2016). Submesoscales play an important role in re-stratifying the mixed layer (e.g. Haine and Marshall 1998; Lapeyre et al. 2006; Boccaletti et al. 2007; Mahadevan et al. 2010, 2012) and enhancing the exchange of water between the mixed layer and pycnocline (e.g. Lévy et al. 2001; Klein and Lapeyre 2009; Thomsen et al. 2016). In addition, submesoscales modify the energetics and fluxes associated with small-scale turbulence in the mixed layer (e.g. D'Asaro et al. 2011; Smith et al. 2016; Taylor 2016). For example, submesoscales transfer energy from large scale geostrophic gradients to small-scale turbulence, while submesoscale stratification in the mixed layer locally inhibits turbulence.

Many submesoscale features are spawned from instabilities associated with horizontal density gradients, or fronts (e.g. Haine and Marshall 1998; Boccaletti et al. 2007; Callies et al. 2016). These instabilities can be interpreted via stability analysis of an "Eady-like" baroclinic zone with parameters characteristic of the mixed layer (e.g. Stone 1966; Stamper and Taylor 2016). Depending on the gradient Richardson number, Ri_g, associated with the vertically-sheared balanced flow, the fastest growing mode is one of two types: mixed layer baroclinic instability (MLI, when $Ri_g > 0.95$) or symmetric instability (SI, when $Ri_g < 0.95$). The most unstable normal mode of MLI is invariant in the cross-front direction and converts available potential energy associated with tilting isopycnals into kinetic energy and ultimately submesoscale eddies, while SI is invariant in the along-front direction and draws its energy from the vertical shear. The net effect of both instabilities is to lower the center of mass of the fluid and increase the stable stratification in the mixed layer. However, submesoscales in the real ocean are a chaotic, nonlinearly-interacting continuum rather than a discrete set of linear modes (e.g. Shcherbina et al. 2013).

Many of the numerical simulations upon which our understanding of non-linear/turbulent submesoscale dynamics is based have either been unforced initial value problems (e.g. Özgökmen et al. 2011; Skyllingstad and Samelson 2012; Stamper and Taylor 2016) or forced with steady surface cooling or winds (e.g. Taylor and Ferrari 2010; Thomas et al. 2013; Hamlington et al. 2014; Taylor 2016). One exception is a study of a storm event at the Gulf Stream front using observations and large-eddy simulations (LES) reported in Thomas et al. (2015). They found turbulent dissipation rates in excess of anticipated values and rapid re-stratification of the boundary layer, and attributed these features to SI. Although they captured SI, the simulations in Thomas et al. (2015) had a limited domain size which excluded the possibility of MLI and hence submesoscale eddies.[1]

[1]Skyllingstad et al. (2017), which was accepted for publication after the submission of this paper, present

Despite the attention paid to submesoscales in recent years, the response of submesoscale eddies to storms is not well understood. Basic open questions remain, including: Can MLI maintain a stable stratification during intense storms? Are submesoscale eddies damped by small-scale turbulent mixing associated with strong winds? How is the small-scale turbulence in the mixed layer influenced by submesoscales during storms?

We address these questions using high-resolution LES, motivated by observations collected near 48.7°N, 16.3°W above the Porcupine Abyssal Plain during the Ocean Surface Mixing, Ocean Sub-mesoscale Interaction Study (OSMOSIS), which reveal significant submesoscale activity throughout the year (Thompson et al. 2016; Buckingham et al. 2016). On September 24-26, 2012, during the deployment cruise, a storm passed over the field site and deepened the mixed layer (Rumyantseva et al. 2015). Glider profiles collected during the storm show that the mixed layer remained well-stratified throughout the storm (their Fig. 4). An idealized representation of this event will be the basis for our analysis.

2. Model description

To elucidate the interaction between submesoscales and small-scale turbulence during the life-cycle of a storm, we present results from a simulation in a large domain that captures the fastest growing MLI length scale, hence the associated energy source for submesoscale eddies, while simultaneously resolving small-scale turbulence. The domain is 1970 m by 1970 m by 80 m covered by a grid with 1024 by 1024 by 160 points that achieves a uniform resolution of 1.9 m by 1.9 m by 0.5 m in x and y and z, respectively. As in Taylor and Ferrari (2010) and Taylor (2016), the flow is expressed as a periodic (in x and y) perturbation from a fixed/constant mean horizontal density gradient $\langle M^2 \rangle_{x,y} = \langle \frac{g}{\rho_0} \frac{\partial \rho}{\partial y} \rangle_{x,y} = 5 \times 10^{-8}$ s^{-2} and

several large eddy simulations of wind forced fronts, expanding on Thomas et al. (2015). However, the analysis also focuses on domains that are too small to permit MLI.

thermal wind shear $\langle M^2\rangle_{x,y}/f = 5\times10^{-4}$ s^{-1} that are representative of the OSMOSIS site before the storm (Christian Buckingham, personal communication). Here, ρ is the density, $\rho_0 = 1026$ kg/m^3 is the reference density, g is the acceleration due to gravity, the Coriolis frequency $f = 10^{-4}$ s^{-1}, and $\langle\rangle_{x,y}$ denotes a horizontal average.

The turbulent state at the onset of the storm (Figure 1 (A)) is obtained from a 3-day spin-up simulation (Whitt 2017) that is forced by a constant air-sea (i.e. surface) buoyancy flux $B_A = 3\times10^{-9}$ m^2/s^3 (buoyancy $b = -g\rho/\rho_0$ is simulated, but this is roughly equivalent to a heat loss of 10 W/m^2 to the atmosphere) and initialized with low-amplitude red noise on a vertical density profile based on Figure 3 (B) of Rumyantseva et al. (2015). The mixed layer depth H_{ML}, which is defined by an increase in the mean density $\langle\rho\rangle_{x,y}$ by .03 kg/m^3 relative to the surface, is initially 35 m. The mixed layer is stratified: $\langle N^2\rangle_{x,y} = \langle -\frac{g}{\rho_0}\frac{\partial\rho}{\partial z}\rangle_{x,y} = 2.5\times10^{-7}$ s^{-2} and the initial balanced Richardson number $Ri_B = f^2\langle N^2\rangle_{x,y}/\langle M^2\rangle_{x,y}^2 = 1$. In the pycnocline, $\langle N^2\rangle_{x,y} = 3.5\times10^{-4}$ s^{-2} and $Ri_B = 1400$. The fastest growing MLI mode has a horizontal scale $L_{\mathrm{MLI}} = \frac{2\pi\langle M^2\rangle_{x,y}H_{ML}}{f}\sqrt{\frac{2+2Ri_B}{5}} \approx 985$ m (Stone 1966), which is half the domain size. The growth timescale of this mode is $T_{\mathrm{MLI}} = \frac{3.3}{|f|}\sqrt{Ri_B+1} \approx 13$ hours.

The storm forcing during September 24-26, 2012 at the OSMOSIS site is represented by the idealized spatially-uniform but time-dependent surface stress in Figure 1 (B), which points 45° to the right of the mean geostrophic flow at the surface. Following the storm, the simulations continue for about 4 days without wind stress to elucidate the subsequent adjustment and re-stratification. In order to separate the effects of storm winds from storm buoyancy fluxes, the air-sea buoyancy flux is held constant at $B_A = 3\times10^{-9}$ m^2/s^3 during and after the simulated storm; this B_A is about ten times weaker than the buoyancy flux associated with the observed cooling during the storm (Rumyantseva et al. 2015).

In order to separate the influence of the front and submesoscales from the classic "one-dimensional" effects of the wind stress on the small-scale turbulence and stratification, the wind-forced simulation in the large domain is compared to a simulation in a small domain without a front or submesoscales. The small domain is 492.5 m by 492.5 m by 80 m and has the same grid resolution, the same surface boundary conditions, and the same mean density profile $\langle \rho \rangle_{x,y}(z)$ at day 0 as the large domain, but $\langle M^2 \rangle_{x,y} = 0$.

In order to identify how the wind modifies the submesoscales, two additional simulations are carried out in the large domain with $\langle M^2 \rangle_{x,y} = 5 \times 10^{-8}$ s^{-2}. These simulations are identical to the baseline simulation described above, except that they are forced only by an air-sea buoyancy flux and the surface stress is zero. In the first of the additional simulations, the buoyancy flux $B_A = 3 \times 10^{-9}$ m^2/s^3 is weak and constant as in the wind-forced simulation. In the second simulation, the buoyancy flux is strong and time dependent; it takes the same magnitude as the Ekman buoyancy flux in the wind-forced simulation, that is $B_A = \mathrm{EBF} = \frac{\tau_x \langle M^2 \rangle_{x,y}}{\rho_0 f}$ (see Figure 1 (B)). Prior work has suggested that the relative strength of the competing de-stratifying Ekman and air-sea buoyancy fluxes and re-stratifying submesoscale buoyancy flux can be quantified using the mixed-layer buoyancy flux ratio:

$$R_{ML} = \frac{B_A + \mathrm{EBF}}{B_{\mathrm{MLI}}}, \tag{1}$$

where the submesoscale buoyancy flux $B_{\mathrm{MLI}} = 2.1 \times 10^{-9}$ m^2/s^3 is a constant derived from a parameterization of MLI assuming a constant mixed layer depth of 37.5 m (Fox-Kemper et al. 2008; Mahadevan et al. 2010, 2012). Here, R_{ML} is between 10-100 during the storm and $R_{ML} = 1.4$ before and after the storm (Figure 1 (B)). Both the wind and strong-buoyancy-flux-forced fronts have the same R_{ML}.

All simulations are carried out with DIABLO (Taylor 2008), which solves the discrete incompressible Boussinesq equations using a pseudospectral method for horizontal derivatives and second-order finite differences for vertical derivatives. Time stepping is accomplished using a third-order Runge-Kutta scheme for advection and the implicit Crank-Nicholson scheme for viscosity/diffusion. The LES solves a filtered version of the governing equations, which are closed using a modified Smagorinsky model to represent sub-grid-scale stresses

(Kaltenbach et al. 1994). The sub-grid-scale diffusivity $\kappa_{SGS} = \nu_{SGS} Pr_{SGS}^{-1}$ depends on the sub-grid-scale viscosity ν_{SGS} and the sub-grid-scale Prandtl number, which is parameterized in terms of the gradient Richardson number at the grid scale $Ri_{GS} = \frac{-g}{\rho_0} \frac{\Delta\rho\Delta z}{\Delta u^2 + \Delta v^2}$, that is $Pr_{SGS}^{-1} = 1/(1 + Ri_{GS}/.94)^{1.5}$ (as in Anderson 2009), where u, v are the horizontal velocities and Δ indicates the difference between two vertically-adjacent grid cells.

3. Results

At the onset of the storm, the density variance in the mixed layer of the large domain is dominated by submesoscales, although the domain contains variability at all resolved scales (Figure 1 (A)). In addition, submesoscale density variability remains a dominant feature of the mixed layer both during and after the storm. The following sections describe the simulated evolution of the mean stratification and small-scale turbulence as well as submesoscale variability within the mixed layer during and after the storm.

a. Mean stratification, shear and dissipation

Both during and after the storm, the mixed layer is characterized by a stronger mean stratification $\langle N^2 \rangle_{x,y}$ and a higher gradient Richardson number, $Ri_g = \langle N^2 \rangle_{x,y} / (\langle \partial u/\partial z \rangle_{x,y}^2 + \langle \partial v/\partial z \rangle_{x,y}^2)$, in the wind-forced front than in the wind-forced domain without a front or the strong-buoyancy-flux-forced front (Figure 2). The stronger stratification implies a higher balanced Richardson number $Ri_B = f^2 \langle N^2 \rangle_{x,y} / \langle M^2 \rangle^2$, which indicates the mean balanced flow is more stable to some classes of instability; $Ri_B > 1$ indicates symmetric stability, and $Ri_B > 0$ indicates gravitational stability. In both simulations with a front, the mean state is stable to gravitational instability ($Ri_B > 0$) and Kelvin-Helmholtz instability ($Ri_g > 1/4$) throughout much of the the mixed layer, despite strong surface momentum or buoyancy fluxes, in contrast to the wind-forced domain without a front.

Despite the strong mean stratification and higher Ri_B throughout much of the mixed layer, the mixing layer depth H_{XL}, where the dissipation rate $\langle\epsilon\rangle_{x,y} > 10^{-8}$ W/kg, is deeper during the storm in the wind-forced front compared to the wind-forced domain without a front or the front forced by a strong air-sea buoyancy flux. In addition, H_{XL} remains deeper than H_{ML} for 0.5 days after the storm is over in the wind-forced front, unlike the other two strongly-forced simulations (Figure 2).

b. *Spatial variability*

The combination of a strongly-stratified and turbulent mixing layer is paradoxical, but it can be explained by spatial variability associated with submesoscales. Both during and after the storm, the stratification (N^2) in the wind-forced front exhibits submesoscale variations of 1-2 orders of magnitude within the mixed layer at all depths (Figures 3 (A) and 4 (A)). Regions of high stratification $N^2 \gtrsim 10^{-5}$ s^{-2}, which dominate the horizontal average, are associated with high potential vorticity, which is much greater than 0, but regions of low-stratification are associated with negative potential vorticity (not shown). Hence, the criteria for SI is met locally in some regions of the domain (Hoskins 1974), but the high mixed-layer stratification cannot be explained by SI, which tends to restore unstable regions with potential vorticity of the opposite sign of f toward conditions neutral to SI with zero potential vorticity and $Ri_B \simeq 1$ (e.g. Taylor and Ferrari 2010; Thomas et al. 2013, 2015). This contrasts with the wind-forced front presented here, where $Ri_B \sim 10$ to 100 in the mixed layer during the storm (Figure 2 (A)), much larger than the neutral state for SI.

During the storm, the submesoscale variability lacks clear coherent vortical structures, but as the storm subsides, a coherent submesoscale cyclonic vortex quickly develops and can be seen by day 3.0 (snapshots at day 3.3 are shown in Figure 4 (A)). This vortex, which has a strongly stratified core and weakly stratified edges, qualitatively dominates the submesoscale variability after the storm (Figure 4 (A)). The vortex diameter is quantitatively consistent with the fastest growing MLI length scale (about 1 km), and it emerges on a time-scale that is quantitatively consistent with the fastest growing MLI timescale (about half a day). However, the vortex forms during the storm, and its growth may be significantly modified by the wind and the associated ageostrophic shear.

The small scale (< 150m) turbulent kinetic energy exhibits spatial variations of 1-2 orders of magnitude within the mixed layer during and after the storm, and the pattern of variability of small scale turbulence is qualitatively similar to the variability in stratification. As a result, strong turbulence penetrates to the mixed layer base in only a small fraction of the domain. Yet, this variability is sufficient to explain why the mixing layer depth H_{XL}, defined using $\langle\epsilon\rangle_{x,y}$ in Figure 2 (A), penetrates deeply into the region of strong mean stratification. The cause of these deep penetrating events is not known, but could be due to local interactions between the wind and the submesoscale fronts and filaments.

c. *Energetics*

The contributions of submesoscales and small scale turbulence to the kinetic energy can be isolated using energy spectra. Here, we focus on the lower part of the mixed layer by presenting spectra at 30 m depth (about 3/4 of H_{ML} after the storm, see Figure 2 (A)). At this depth, the horizontal and vertical kinetic energy spectra have different slopes at large and small scales (Figure 5 (A)-(B)). In addition, the vertical kinetic energy spectra exhibit two local maxima, one at a wavenumber of about 1/1000 cycles/m (near the fastest growing MLI mode), and one at a wavenumber between 1/50 and 1/100 cycles/m. This motivates using a cutoff wavenumber $k_c = 1/150$ cycles/m, near the local minimum in the vertical kinetic energy spectra (see Figure 5 (B)), to separate large from small scales.

Large-scale horizontal kinetic energy dominates the total kinetic energy in the wind-forced front. It grows during the storm and decays to about 25% of its late-storm maximum after the end of the storm (Figure 5 (C)). In contrast, large-scale horizontal kinetic energy rises only slightly in the front forced by a weak air-sea buoyancy flux and decays during forcing in the front forced by a strong air-sea buoyancy flux. Hence, the total kinetic energy is more than ten times larger during the storm in the wind-forced front than in any of the other three simulations (Figure 5 (C)-(D)).

Large-scale vertical kinetic energy is about ten times larger during the storm than before or after the storm in the simulation with a wind-forced front (Figure 5 (D)), which is qualitatively consistent with earlier studies that show wind enhances submesoscale vertical motions at fronts (e.g. Mahadevan and Tandon 2006; Thomas et al. 2008). However, the large-scale vertical kinetic energy is also enhanced during the storm in the simulation forced by a strong buoyancy flux (Figure 5 (D), dashed red line). Comparing Figures 5 (B) and (D), it is evident that the large-scales are highly anisotropic at a wind-forced front (blue lines), while strong convective forcing (red lines) causes the flow to become more isotropic (although the large-scale horizontal kinetic energy is still more than 10 times larger than the vertical kinetic energy in this case.)

During the storm, the small-scale turbulent kinetic energy is similar in all three simulations with strong surface forcing (Figure 5). However, after the storm, small-scale turbulence is less energetic and small-scale spectral slopes are steeper for the wind-forced front compared to the simulation without the front (Figure 5), presumably because the submesoscale re-stratification suppresses small-scale turbulence at 30 m depth in the simulation with the front (see Figure 4). Yet, small scale turbulence is more energetic in the large domain during a transition period just after the storm, e.g. between days 2.75 and 4 (Figures 5 (C)-(D)), which explains why H_{XL} remains deeper than H_{ML} after the storm (Figure 2 (A)) and suggests that mixing can decouple (in time) from wind-forcing at fronts (as in Whitt et al.

2017).

4. Conclusion

It has been known for some time that submesoscales can have a significant impact on stratification and small-scale turbulence in the ocean mixed layer. This work expands our understanding of submesoscale dynamics by presenting high-resolution large eddy simulations that elucidate the interaction between submesoscales and small-scale turbulence during the life-cycle of a mid-latitude storm. We find that submesoscales persist and even grow during strong winds. Contrary to existing theory and simulation results (Mahadevan et al. 2010), which suggest that submesoscale re-stratification should be overwhelmed by the destratifying effects of the Ekman buoyancy flux, our simulations show that submesoscales maintain strong mean stratification in the mixed layer even in the midst of strong downfront winds. Despite the strong mean stratification, small-scale turbulence intermittently penetrates to the mixed layer base due to strong modulation of mixed-layer stratification on submesoscales. The small-scale turbulent kinetic energy is enhanced in regions of relatively weak stratification, both during and after the storm.

The persistence of strong, stable stratification during the storm, first reported by Rumyantseva et al. (2015), and confirmed here by the LES, challenges the prevailing description of submesoscales. Recent work has framed the description of the mixed layer depth and stratification as a competition between re-stratification by submesoscales associated with horizontal density gradients and mixing by small-scale turbulence associated with surface forcing (e.g. Mahadevan et al. (2010, 2012); Bachman and Taylor (2016); Taylor (2016)). The results here suggest a more nuanced description where winds simultaneously energize small-scale turbulence *and* submesoscales. Notably, the submesoscale horizontal kinetic energy is significantly enhanced during the storm (see Figure 5 (B)). Despite the enhanced small-scale turbulence and the large de-stabilizing Ekman buoyancy flux and large values of the mixed-layer buoyancy flux ratio (R_{ML}), strong stratification persists in localized patches (Figure 3 (A)). The same level of stratification is not seen in a simulation with the same R_{ML} without wind forcing, suggesting that the enhancement of submesoscale activity by wind forcing is important for the evolution of mixed layer stratification.

These results raise several important questions for future work, including: Is MLI enhanced by small-scale ($<$ 1 km) buoyancy gradients and/or strong Ekman shear? Does the domain size constrain the dynamics of the submesoscales? Do surface waves, which are excluded here, modify the results? Finally, how do the results depend on the chosen parameters, including the horizontal and vertical density gradients, the wind stress, and the air-sea buoyancy flux?

Although only one set of parameters is considered here, this set of parameters is typical of the OSMOSIS site (Thompson et al. 2016) and presumably is relevant to other regions of the ocean. Moreover, the simulated strong stratification during the storm is qualitatively consistent with the observed mixed layer stratification at the OSMOSIS site during the September storm (Rumyantseva et al. 2015). Hence, the results, which challenge our current understanding of submesoscale dynamics, could provide insight into typical ocean conditions during the passage of a storm.

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List of Figures

5 Time-averaged power spectra of horizontal velocity E_h (A) and vertical velocity E_v (B) at $z = -30$ m as a function of radial horizontal wavenumber $|k_h|$. Time series of horizontal kinetic energy $\mathcal{E}_h = \int E_h dk_h$ (C) and vertical kinetic energy $\mathcal{E}_v = \int E_v dk_h$ (D) in the wind-forced front (blue), the strong-buoyancy-flux-forced front (red), the weak-buoyancy-flux-forced front (gray), and the wind-forced domain without a front (green). The wavenumber spectra in (A)-(B) are averaged during the storm ($0.5 < t < 2.75$ days, solid) and after the storm ($4.5 < t < 7$ days, dashed lines). The kinetic energy in (C)-(D) is integrated over small scales (dotted), that is over wavenumbers $|k_h| > k_c$ where $k_c = 1/150$ cycles/m, and large scales (dash-dotted), that is $|k_h| < k_c$. Several lines are omitted: solid gray lines are omitted from (A)-(B) and dotted gray lines are omitted from (C)-(D) because there is no storm event in that simulation. Dashed red lines are omitted from (A)-(B) because the simulation is not run for the post-storm period. Finally, dash-dotted green lines are omitted from (C)-(D) because the magnitude is low. 25

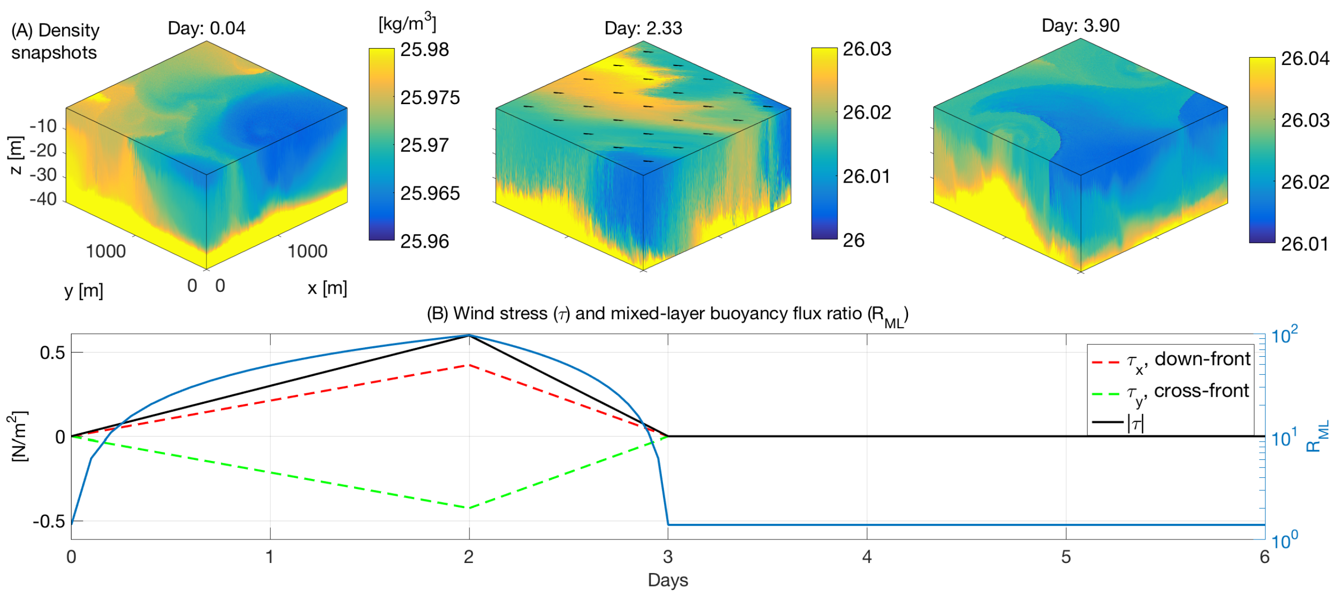

FIG. 1. (A) Snapshots of density and (B) time series of wind stress magnitude (black) and vector components (dashed red and green) as well as the mixed layer buoyancy flux ratio R_{ML} (blue) [see (1)]. Black vectors in the snapshot at day 2.33 indicate the direction of the wind during the storm.

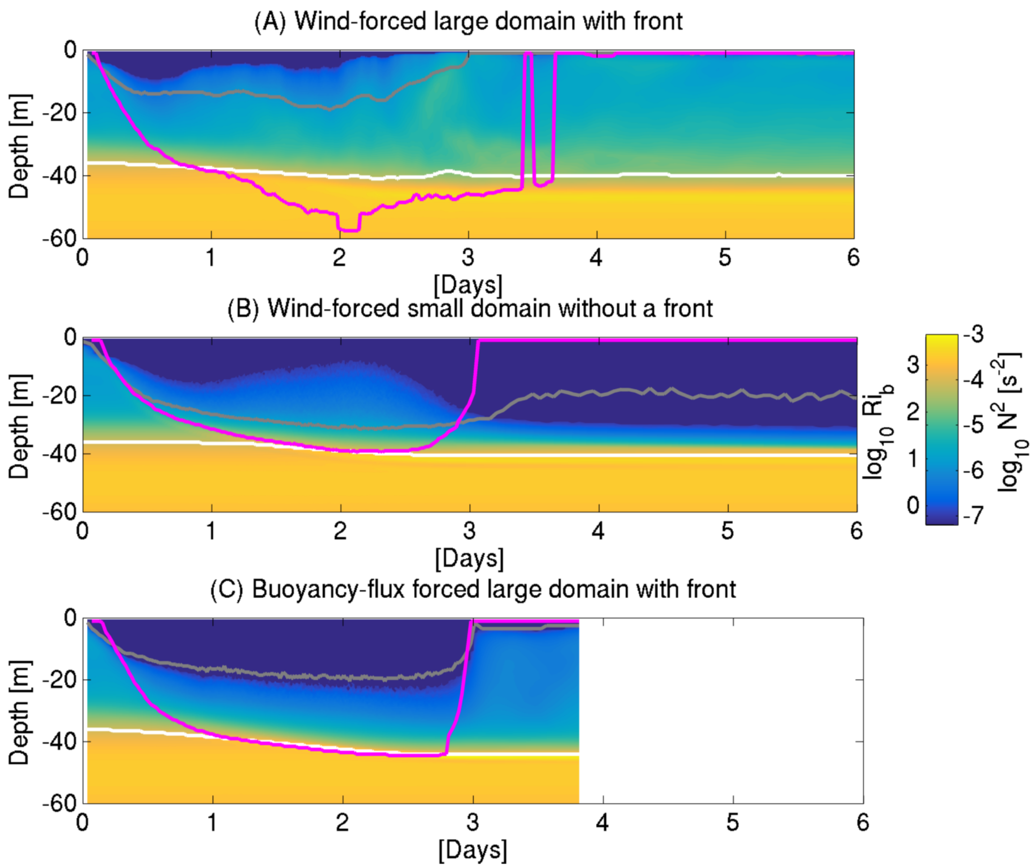

FIG. 2. Time series of horizontally-averaged stratification $\langle N^2\rangle_{x,y}$ and (equivalently) the balanced Richardson number $Ri_B = f^2\langle N^2\rangle_{x,y}/\langle M^2\rangle^2_{x,y}$ in three simulations: the wind-forced front (A), the wind-forced domain without a front (B), and the strong-buoyancy-flux-forced front (C). Panels also include time series of mixed layer depth H_{ML} (white), mixing layer depth H_{XL} (magenta), and the low-gradient-Richardson-number depth H_{Ri} (gray), above which the gradient Richardson number $Ri_g \leq 1/4$.

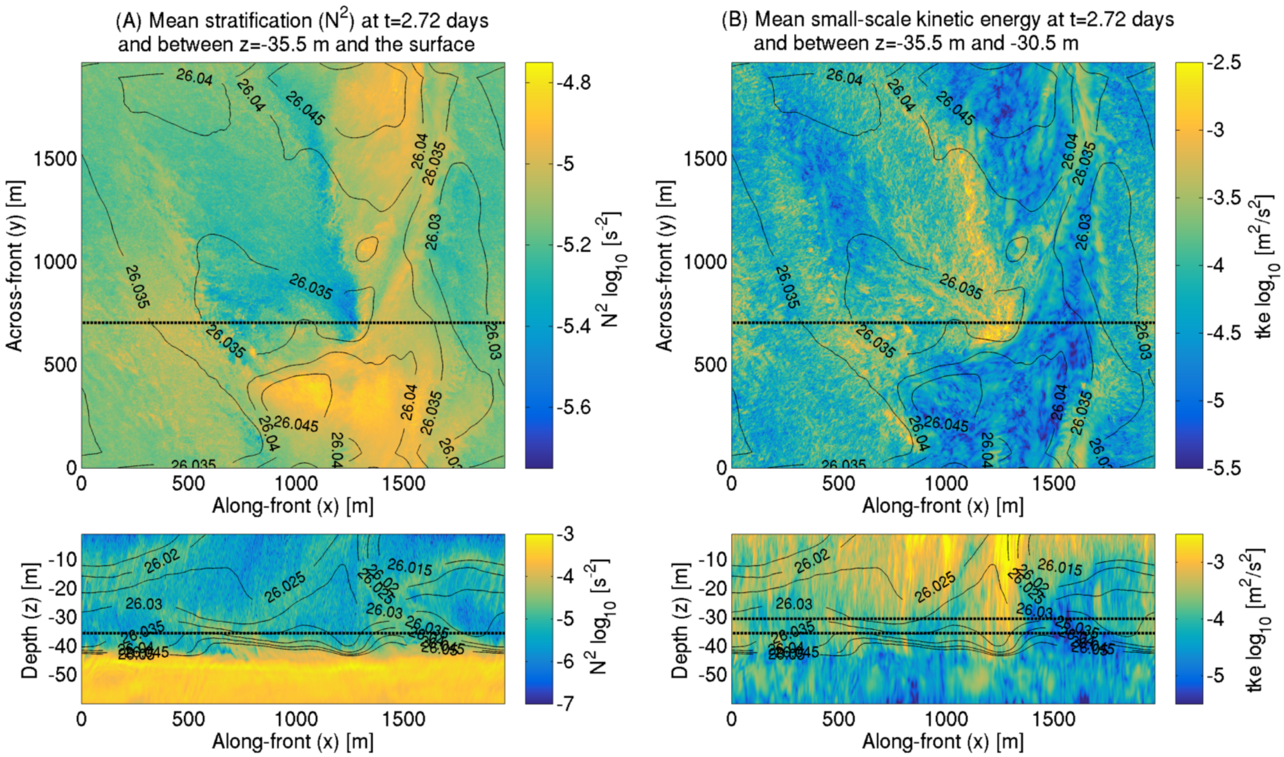

FIG. 3. Snapshots of (A) stratification N^2 and (B) small-scale turbulent kinetic energy at t=2.72 days in the wind-forced front (just before the end of the storm, see Figure 1 (B)). Solid black contours of the large-scale density are overlaid. The x-y slices (top) are calculated as an average from $z = -35.5$ m to the surface in (A) and to -30.5 m in (B). The x-z slices (bottom) are calculated at the y location indicated by the dashed black lines in the x-y slices. Here, large scales are defined by applying a 150 m by 150 m square filter to the full fields at each vertical level, while small scales are defined as the difference between the full fields and the large scale fields.

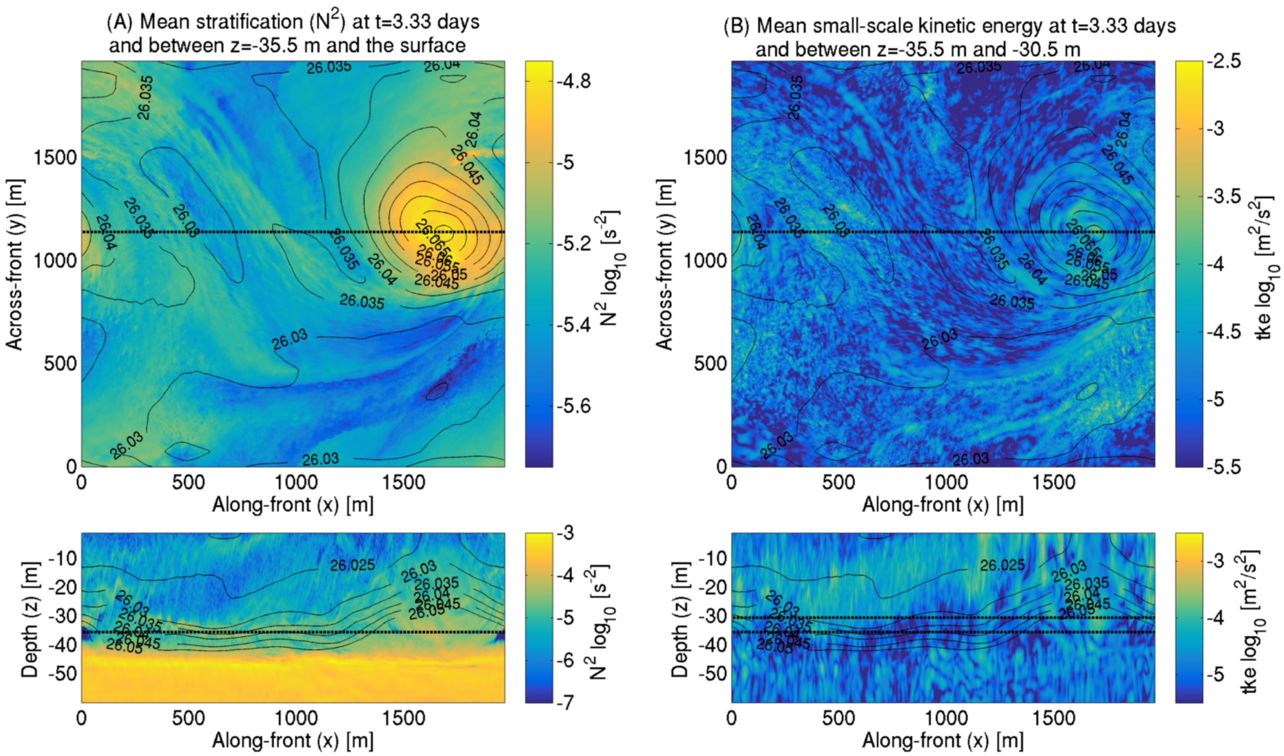

FIG. 4. As in Figure 3, but at t=3.33 days, just after the storm is over; see Figure 1 (B).

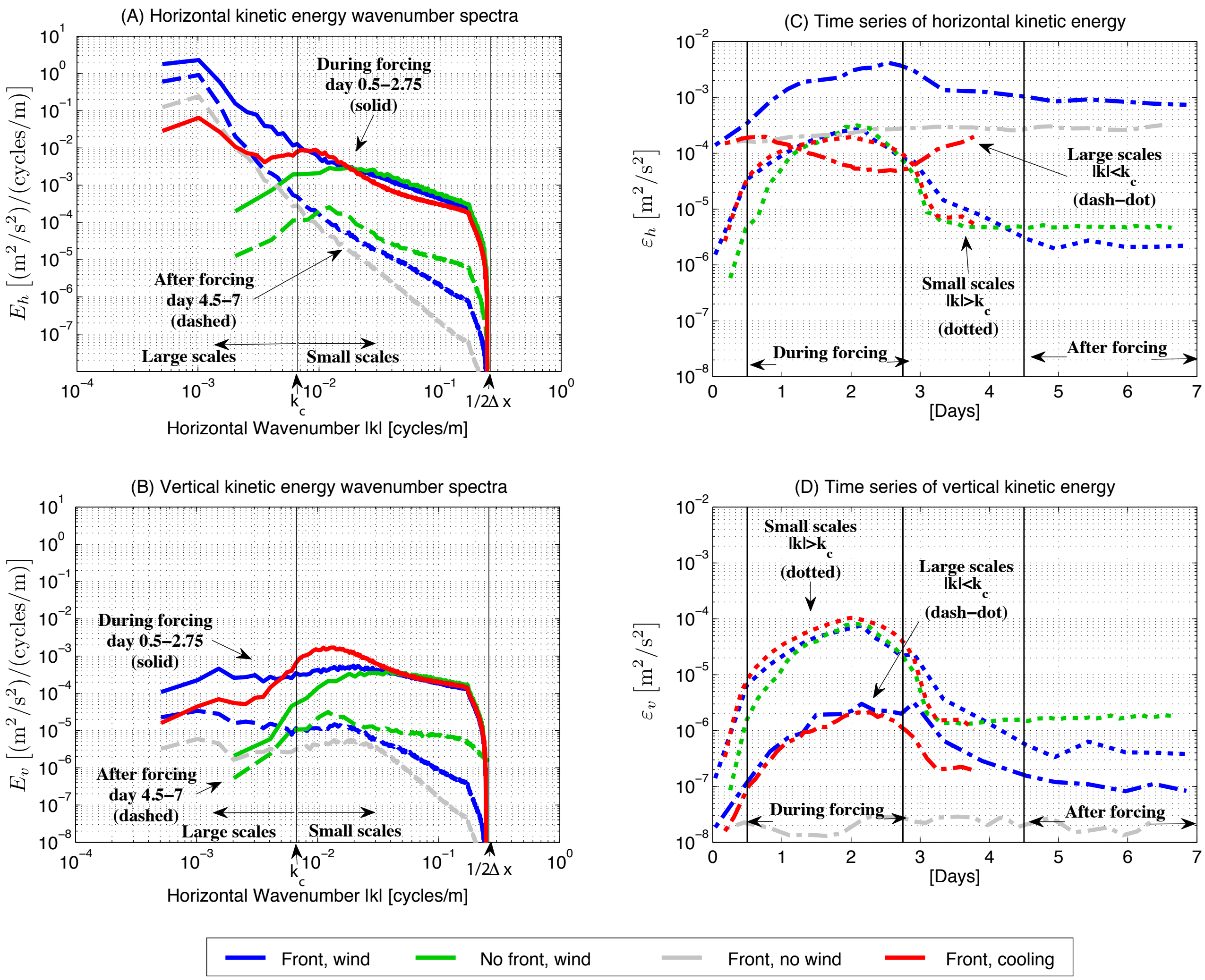

FIG. 5. Time-averaged power spectra of horizontal velocity E_h (A) and vertical velocity E_v (B) at $z = -30$ m as a function of radial horizontal wavenumber $|k_h|$. Time series of horizontal kinetic energy $\mathcal{E}_h = \int E_h dk_h$ (C) and vertical kinetic energy $\mathcal{E}_v = \int E_v dk_h$ (D) in the wind-forced front (blue), the strong-buoyancy-flux-forced front (red), the weak-buoyancy-flux-forced front (gray), and the wind-forced domain without a front (green). The wavenumber spectra in (A)-(B) are averaged during the storm ($0.5 < t < 2.75$ days, solid) and after the storm ($4.5 < t < 7$ days, dashed lines). The kinetic energy in (C)-(D) is integrated over small scales (dotted), that is over wavenumbers $|k_h| > k_c$ where $k_c = 1/150$ cycles/m, and large scales (dash-dotted), that is $|k_h| < k_c$. Several lines are omitted: solid gray lines are omitted from (A)-(B) and dotted gray lines are omitted from (C)-(D) because there is no storm event in that simulation. Dashed red lines are omitted from (A)-(B) because the simulation is not run for the post-storm period. Finally, dash-dotted green lines are omitted from (C)-(D) because the magnitude is low.