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Coupling turbulence, surface waves, and submesoscale cold filaments in the upper ocean boundary layer using LES

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The upper ocean is a high-Reynolds number $\mathcal{O}(10^8)$ turbulent flow featuring an abundance of dynamical processes that generate motions over a broadband spectrum of time and space scales. For example, extensive work now demonstrates an active and prevalent "submesoscale turbulence" regime full of fronts, density filaments, and coherent vortices with lateral dimensions of 100 m to 10 km [1]. At the same time, ocean boundary layer (OBL) turbulence is modified by organized streamwise rolls generated by surface gravity waves, so called "Langmuir turbulence", that extend over horizontal distances of a kilometer even when the OBL is only tens of meters deep [2]. The present work seeks to combine submesoscale and Langmuir turbulence. In the OBL, surface waves are nearly irrotational and propagate rapidly compared to the underlying rotational slow-moving currents. This scale separation allows for elegant asymptotic methods to be applied with time-averaging carried out over the rapidly oscillating waves, *i.e.,* wave-phase averaging. This asymptotic theory was first derived by Craik & Leibovich [3] and later extended by McWilliams *et al.* [4, 5]. In a large-eddy simulation (LES) model of the OBL that adopts the Craik-Leibovich-McWilliams asymptotics the water surface is flat and wave effects appear as Stokes advection in all scalar equations and as a Stokes-Coriolis term and generalized vortex force,

$$\frac{\partial \overline{\boldsymbol{u}}}{\partial t} \;=\; \ldots \;+\; \mathbf{u}^{St} \times (f\hat{\mathbf{z}} + \overline{\boldsymbol{\omega}}) \;+\; f\hat{\mathbf{z}} \times \boldsymbol{u}_g \;,$$

in the resolved scale momentum equations: f is the Coriolis parameter, the resolved vorticity $\overline{\boldsymbol{\omega}} = \nabla \times \overline{\boldsymbol{u}}$ and $\mathbf{u}^{St}$ is the Lagrangian-Stokes drift velocity induced by the wave field. The new feature included in the present work is forcing by large-scale geostrophic currents $\boldsymbol{u}_g$. These currents are assumed to satisfy the thermal-wind balance relationships [8] and are chosen here as idealizations of observed submesoscale cold filaments [9]. For an externally imposed two-dimensional large-scale temperature field (a filament) $\Phi(x, z)$ the consistent horizontal pressure gradients which appear in the momentum equations are then embodied in the large-scale geostrophic currents $\boldsymbol{u}_g = \boldsymbol{u}_g(x, z)$.

To investigate the coupling between turbulence, surface waves, and submesoscale temperature fronts we use large-domain fine-resolution LES. Our LES code solves incompressible Boussinesq equations using pseudospectral/finite-difference methods and employs custom matrix transposes to achieve 2D domain decomposition [10]. This algorithm allows a large number of computational cores to be utilized even in anisotropic physical domains that are relatively thin in their vertical dimension compared to their horizontal extent.

A recent application of our LES seeks to unravel what time-varying combinations of wind, wave and cooling at the water surface lead to interior mixing and deep mixed layers in the Southern Ocean. To this end, a suite of simulations are carried out for transient wind-wave regimes with rapidly increasing/decreasing winds designed to produce sea states with growing and decaying swell waves. This allows us to examine OBL mixing sensitivity and the coherent structures in an environment with rapidly changing values of the turbulent Langmuir number La_t [4]. Fig. 1 is an illustration of how the near surface maximum in vertical velocity varies with time for different wind spin-down rates; for each simulation the starting condition is steady wind-wave equilibrium and the simulations use a time varying vertical profile of Stokes drift derived from the large-scale wave prediction code WaveWatch III [6]. As expected, the simulations produce a normalized maximum $w^2_{max}/u^2_* = 3$ in the steady regime $t < 25$ hrs indicating that Langmuir turbulence is active and dominant. Each simulation however apparently follows a unique trajectory on its path to a swell dominated regime depending on the wind spin-down rate. We were further surprised at the steady decline of the maximum vertical velocity for $t > 40$ hrs suggesting that the magnitude of the Stokes profile under these swell dominated conditions erodes somewhat quickly with time. We intend to further compare these results with OBLs driven by non-equilibrium winds and waves under hurricane conditions [7].

In contrast, Fig. 2 shows a typical view of the vertical velocity from a simulation under wind-wave equilibrium conditions, *viz.,* winds of 20 $\mathrm{m\,s^{-1}}$ with significant wave height $H_s \sim 10$ m. The computations are carried out using a mesh of $(1024 \times 1024 \times 320)$ points in a physical domain $(1000 \times 1000 \times 350)$ m running on 1024 computational cores for more than 300,000 time steps ~ 67 physical hours. The long integration time and the presence of wave-current interactions causes the mixed layer depth h to deepen from 30 m to more than 130 m, a typical value for the Southern Ocean. The wide dimensions of the computational domain permit large-scale vigorous Langmuir cells to develop under the action of the strong persistent surface winds.

A new area of computational research we are just beginning to explore is to examine the two-way interaction between OBL turbulence and submesoscale temperature fronts. This is a challenging computational problem as it requires fine meshes to resolve OBL turbulence, but in computational domains large enough to support realistic large-scale temperature fronts. Results from one of our recent simulations are shown in Figs. 3 and 4. In this example, we solve the LES equations in a domain $(18000 \times 562.5 \times 250)$ m using a mesh of $(12288 \times 384 \times 256)$ gridpoints, *i.e.,* with horizontal spacing $\triangle x = \triangle y = 1.46$ m and vertical spacing $\triangle z = 0.5$ m near the water surface. The nominal width of the cold temperature front is 6 km (see Fig. 3). A two-step process is used to conduct these calculations. First, we run simulations with no-fronts for more than 60 physical hours to generate equilibrium turbulence. A data volume from these no-front computations is then used to initialize the frontal simulations, which are typically carried forward for an additional 12 physical hours. The simulations beautifully illustrate cold filament frontogenesis with intense sharpening of both the (u, v) currents. One of the intriguing results is shown in Fig. 4. The vertical velocity shows an abrupt change in activity as the frontal boundary is crossed despite the presence of uniform surface winds. We speculate the formation of the secondary circulations (u, w) induced by the temperature field acts in concert with the winds on the left side of the front while they are opposed on the right side of the front - thus effectively reducing the turbulence levels. We also find that the sharpness of the front stabilizes with time under the action of resolved turbulence. These simulations with and without wave effects will provide clues as to the dynamical processes that limit frontogenesis.

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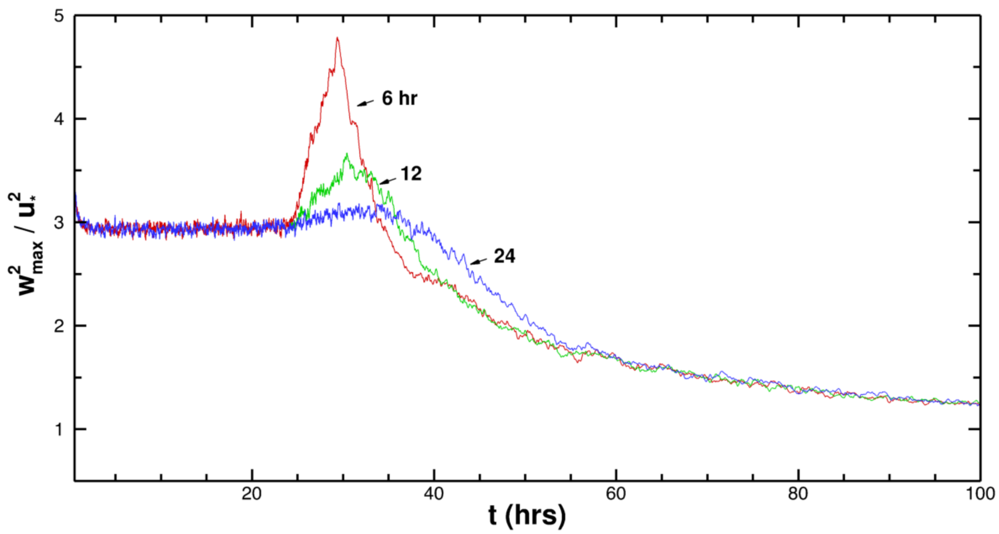

Figure 1: *Temporal variation of the maximum vertical velocity variance w^2_{max} for transient wind-wave regimes. In each simulation the winds spin down at $t \sim 25$ hrs from 20 to 4 $m\,s^{-1}$ but with varying decay periods of $(6, 12, 24)$ hours. The vertical velocity variance is normalized by the local time varying value of the wind stress $u^2_*(t)$.*

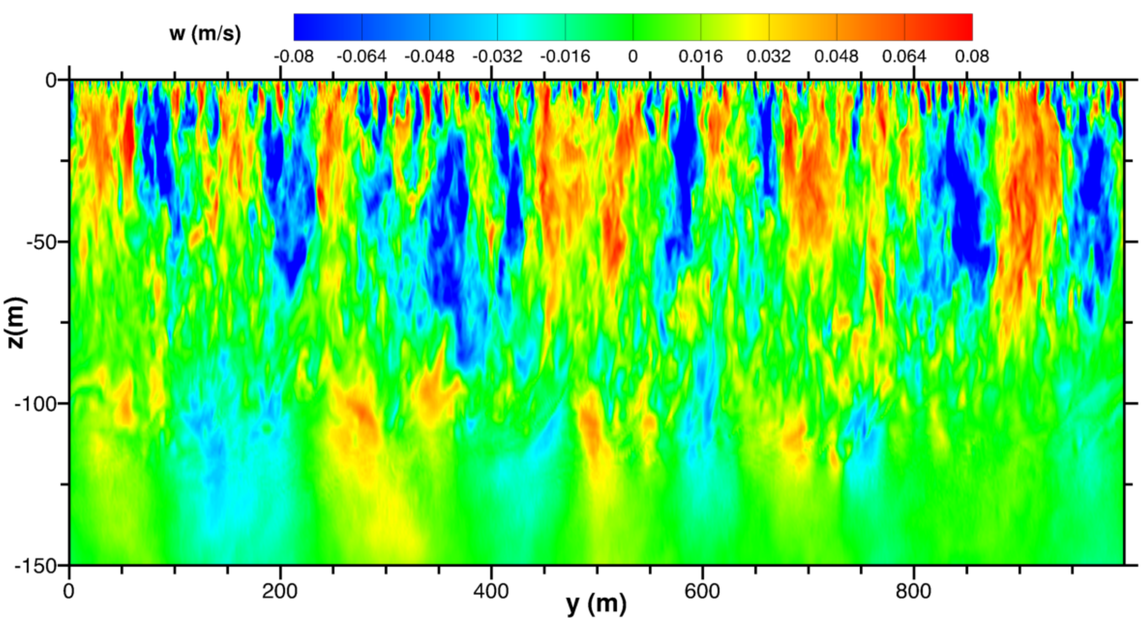

Figure 2: *Contours of vertical velocity w in a y-z plane illustrating the formation of turbulent Langmuir cells in a simulation of the stratified Southern Ocean boundary layer. The depth of the downwelling zones (blue contours) extend to $z \sim -100$ m. The surface winds are 20 $m\,s^{-1}$ and are oriented in the $+\hat{\mathbf{x}}$ direction (out of the page). The Stokes drift, which appears in the LES equation set, is computed from the large scale wave prediction code WaveWatch III [6] assuming wind-wave equilibrium. The significant wave height $H_s \sim 10$ m.*

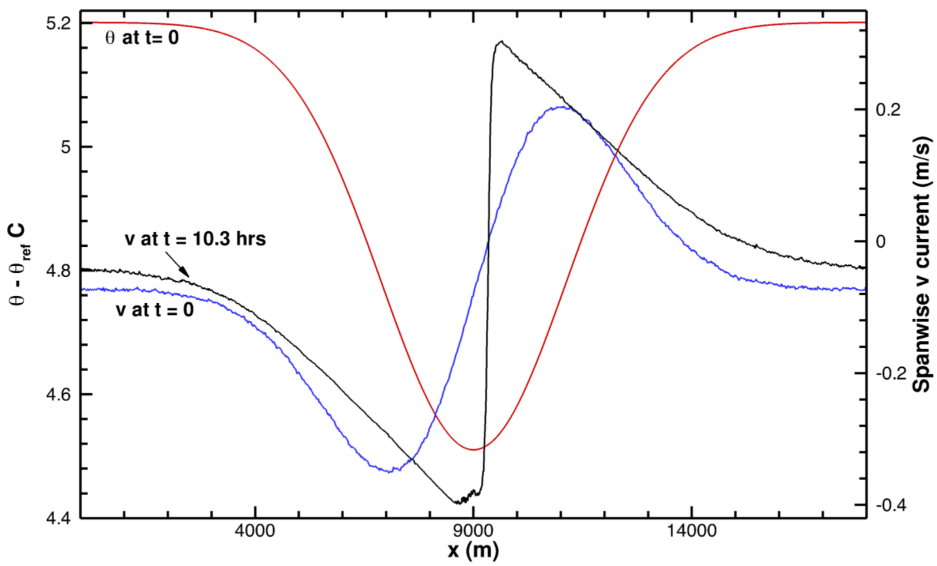

Figure 3: *Variation of spanwise v current near the water surface $z = -0.5$ m in the presence of a cold temperature filament (front) and resolved turbulence. The wind stress τ, corresponding to surface winds $\sim$ 8.2 $m\,s^{-1}$, acts in the west-east $+\hat{\mathbf{x}}$ direction. The θ and v fields are averaged across y. The current profile contains both large scale geostrophic and ageostrophic Ekman boundary layer currents. The extreme sharpening of the current, and its horizontal gradient the vertical vorticity $\zeta = \partial v/\partial x$, illustrate submesoscale frontogenesis [1].*

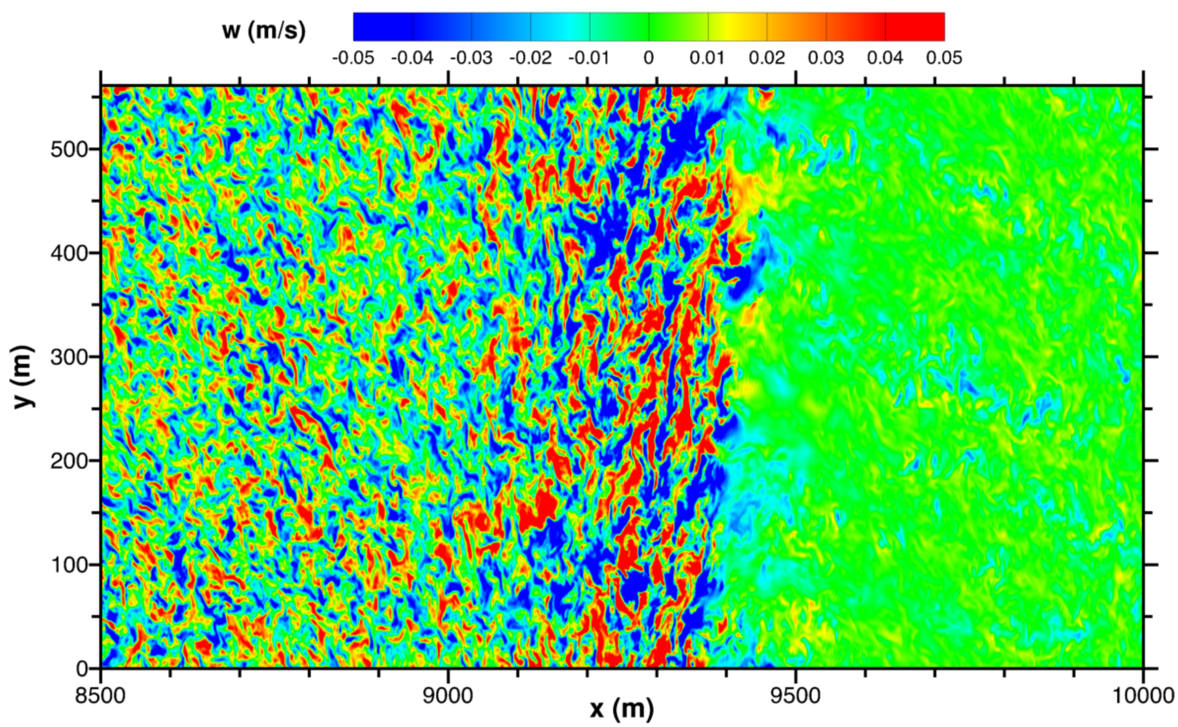

Figure 4: *Contours of vertical velocity w at $z = -60$ m at $t \sim 10.3$ hrs from the LES described in Fig. 3. Notice the difference in activity level between the left and right sides of the front. At this time, the frontal boundary has drifted eastward because of Ekman transport.*