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Parameterization of Frontal Symmetric Instabilities. I: Theory for Resolved Fronts

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A parameterization is proposed for the effects of symmetric instability (SI) on a resolved front. The parameterization is dependent on external forcing by surface buoyancy loss and/or down-front winds, which reduce potential vorticity (PV) and lead to conditions favorable for SI. The parameterization consists of three parts. The first part is a specification for the vertical eddy viscosity, which is derived from a specified ageostrophic circulation resulting from the balance of the Coriolis force and a Reynolds momentum flux (a turbulent Ekman balance), with a previously proposed vertical structure function for the geostrophic shear production. The vertical structure of the eddy viscosity is constructed to extract the mean kinetic energy of the front at a rate consistent with resolved SI. The second part of the parameterization represents a near-surface convective layer whose depth is determined by a previously proposed polynomial equation. The third part of the parameterization represents diffusive tracer mixing through small-scale shear instabilities and SI. The diabatic, vertical component of this diffusivity is set to be proportional to the eddy viscosity using a turbulent Prandtl number, and the along-isopycnal tracer mixing is represented by an anisotropic diffusivity tensor.

Preliminary testing of the parameterization using a set of idealized models shows that the extraction of total energy of the front is consistent with that from SI-resolving LES, while yielding mixed layer stratification, momentum, and potential vorticity profiles that compare favorably to those from an extant boundary layer parameterization (Large et al., 1994). The new parameterization is also shown to improve the vertical mixing of a passive tracer in the LES.

1. Introduction

The oceanic surface mixed layer is the contact point between the atmosphere and the ocean, and is responsible for communicating atmospheric fluxes into the ocean interior, including fluxes associated with heat, momentum, carbon, oxygen, and other tracers. In addition to influencing transport into the ocean interior, atmospheric fluxes strongly influence the dynamics within the mixed layer itself and drive vertical mixing. At the same time, lateral density gradients tend to increase the vertical stratification within the mixed layer (Tandon and Garrett, 1994, 1995; Hosegood et al., 2006). Submesoscale turbulence is the byproduct of the dynamical interplay between lateral density gradients and atmospheric forcing, and has been the focus of many recent studies (e.g. Boccaletti et al., 2007; Capet et al., 2008b–d; Thomas et al., 2008; Thomas and Ferrari, 2008; Mahadevan et al., 2010; McWilliams, 2010; Hamlington et al., 2014; Callies et al., 2016).

The oceanic submesoscale is typically defined as occupying a range of horizontal scales between 100 m and 10 km. Models with an O(1) km horizontal grid permit many of the important dynamical processes at these scales (e.g., Oschlies, 2002), especially the largest fronts and baroclinic instabilities. A more precise definition of submesoscales are those which are marginally constrained by rotation and stratification (the Rossby and Richardson numbers are both $\mathcal{O}(1)$, Thomas et al., 2008) and thus feature both geostrophic and ageostrophic components. Submesoscale dynamics are typically considered to be hydrostatic, which places a lower bound on the range of horizontal scales near the mixed layer depth, or O(100) m, as the aspect ratio approaches unity. Because the Rossby number is near one, the characteristic timescale is $1/f \simeq \mathcal{O}(1)$ day, which is much faster than the mesoscale and results in submesoscales playing a leading role in the evolution of the mixed layer in response to atmospheric forcing. Finally, submesoscale eddies

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frequently arise as a result of straining by mesoscale eddies, and their own strain fields in turn beget submesoscale frontogenesis (Shakespeare and Taylor, 2013, 2014) and a wide variety of small-scale instabilities that cascade energy down to dissipative scales (Capet et al., 2008b; Molemaker and McWilliams, 2010).

It is the fast evolution of the submesoscale in conjunction with atmospheric forcing that is of principal dynamical interest here. The submesoscale features lateral density gradients that arise via eddy straining, preconditioning the flow to convective instabilities driven by both the wind and surface buoyancy forcing. In particular, when the wind blows in a down-front direction the Ekman transport of dense fluid over light can destabilize the mixed layer. Buoyancy forcing can either exacerbate this destabilization in the case of a positive flux (where the positive z direction is taken to be out of the ocean) or mitigate it for a negative flux. Which among the “zoo” of possible submesoscale instabilities that may arise depends heavily on the local relative vorticity (Haine and Marshall, 1998; Thomas et al., 2013; Shcherbina et al., 2013).

Of the many types of submesoscale instability that arise in the mixed layer, baroclinic instability has thus far received the most attention (e.g. Boccaletti et al., 2007; Fox-Kemper et al., 2008). From a modeling perspective, the dynamics of baroclinic instability are fairly well-understood and have led to an extensive body of research about how it should be parameterized in eddy-free general circulation models (GCMs). The success of the most well-known of these parameterizations, the Gent and McWilliams (1990) parameterization for mesoscale baroclinic instability, has led to several subsequent papers regarding the nature of tracer transport by subgridscale motions in a variety of dynamical regimes (Gent et al., 1995; Tandon and Garrett, 1996; Dukowicz and Smith, 1997; Killworth, 1997; Treguier et al., 1997; Griffies, 1998; Griffies et al., 1998; Greatbatch, 1998; Smith and Gent, 2004, and more). A similar conceptualization to that of Gent and McWilliams (1990) underlies the parameterization of mixed-layer, ageostrophic baroclinic instability at submesoscales (Fox-Kemper et al., 2008; Fox-Kemper and Ferrari, 2008; Fox-Kemper et al., 2011).

A new modeling challenge arises when the resolution of an ocean model permits partial resolution of the eddy field. At this point parameterizations must be carefully constructed and previous parameterizations carefully recast so that they do not either outcompete the resolved eddies for energy (e.g. Henning and Vallis, 2004) or double-count the effects of the eddies (Delworth et al., 2012). Unlike Gent and McWilliams (1990), the parameterization of Fox-Kemper et al. (2011) explicitly depends on the model grid scale and thus exemplifies how physical scalings may allow parameterizations to be recast to adapt to increasing model resolution. Fox-Kemper et al. (2011) can be used even when large mixed layer eddies (hereafter MLE) are resolved. The parameterization for *oceanic symmetric instability* (hereafter SI) put forth here is intended to be used in models where some fronts and large-scale frontal instabilities (e.g., baroclinic instabilities) are resolved or handled with scale-adaptive parameterizations, but the smaller frontal instabilities, in particular symmetric instabilities, are not resolved. Many high-resolution nested and regional models fall into this category (e.g., Capet et al., 2008a, 2008b; Lévy et al., 2010; Sutherland et al., 2011; Mensa et al., 2013; Zhong and Bracco, 2013; Molemaker et al., 2015; Rosso et al., 2015; Siedlecki et al., 2015; Gula et al., 2016).

In addition to parameterizations for mesoscale and submesoscale baroclinic instability, many parameterizations exist for small-scale turbulence and mixing. Shear and convective instability are commonly parameterized (for example by Kraus and Turner, 1967; Mellor and Yamada, 1982; Large et al., 1994) and, more recently, Langmuir turbulence (e.g., McWilliams and Sullivan, 2000; Smyth et al., 2002; Grant and Belcher, 2009; Van Roekel et al., 2012; Harcourt, 2013; Li et al., 2016). Recent work has shown that the mixed layer stratification, energy dissipation, and resolved eddy behavior can be highly sensitive to the details of such parameterizations (e.g. Mukherjee et al., 2016). To date, however, no parameterization exists for SI despite observations of its effects on the shear, stratification, and dissipation of kinetic energy in the surface mixed layer (D'Asaro et al., 2011; Thomas et al., 2016).

Numerical simulations with an $\mathcal{O}(1)$ km grid that resolve submesoscale fronts and larger MLE typically do not resolve smaller SI. Thus, it is desirable to have a parameterization for the $\mathcal{O}(100)$ m SI so that submesoscale simulations of an active field of fronts do not require the added cost of resolving the SI (e.g., Fox-Kemper and Ferrari, 2008), just as KPP or other turbulence parameterizations are used to avoid the cost of resolving $\mathcal{O}(1)$ m features (e.g., Hamlington et al., 2014). The goal of this paper is to propose a framework for a parameterization that approximates the restratification and tracer mixing by SI in the case where SI modes are unresolved, but the front that undergoes SI is resolved. The performance of this parameterization under different forcing scenarios is evaluated against the results of SI-resolving Large Eddy Simulations (hereafter LES, e.g. Taylor and Ferrari, 2010; Thomas et al., 2013; Hamlington et al., 2014) in some representative cases.

2. Basics of symmetric instability

The principal focus here, SI, arises in baroclinic flows featuring a lateral density gradient and an associated vertically sheared geostrophic flow. SI is typified by overturning circulations about an axis aligned with the geostrophic flow, typically with the flow along density surfaces (Eliassen, 1949). Assuming a geostrophically balanced flow with buoyancy gradients $N^2 = \partial b/\partial z$ and $\nabla_h b = (\partial b/\partial x, \partial b/\partial y)$, the SI growth timescale T and horizontal lengthscale L may be estimated for constant shear and stratification (Stone, 1966) as

$$\min(T) = \frac{H}{U}\frac{\sqrt{Ri_b}}{\sqrt{1-Ri_b}}, \qquad \max(L) = 2\frac{U}{f}\sqrt{1-Ri_b}, \tag{1}$$

where Ri_b is the balanced Richardson number,

$$Ri_b = \frac{N^2 f^2}{|\nabla_h b|^2}. \tag{2}$$

For typical ocean mixed layer parameters in conditions favorable for SI, $0.01 \le U \le 0.1$ m s^{-1}, $25 \le H \le 100$ m, and $0.25 \le Ri_b \le 0.95$, which imply SI timescales ranging from one minute to an hour and lengthscales of 50–2500 m. These are much smaller than the timescale of 14–18 h and lengthscale of 600 m to 8 km for MLE (e.g. Fox-Kemper et al., 2008, Eq. (2) and (3)), highlighting the potential importance of SI to the evolution of the mixed layer on fast time scales. The corresponding SI mixing rates can be estimated using in situ measurements of density, wind, and surface heat flux, and are compared against those of MLE in Appendix A.

SI occurs when the sign of the Ertel potential vorticity

$$q = \left(f\hat{\mathbf{k}} + \nabla\times\mathbf{u}\right)\cdot\nabla b \tag{3}$$

is anticyclonic – i.e., where q is opposite in sign to the Coriolis parameter f, so that $fq < 0$ (Hoskins, 1974). Here $\mathbf{u}$ is the velocity, $\hat{\mathbf{k}}$ is a unit vector in the vertical, and $b = -g(\rho-\rho_0)/\rho_0$ is the buoyancy, which is defined in terms of the gravitational acceleration g, density ρ, and constant background density ρ_0. The condition $fq < 0$ is most straightforwardly satisfied in convective conditions ($N^2 < 0$) which can give rise to gravitational instability, or inertial instability when $N^2 > 0$ and $\zeta_a < 0$, where $\zeta_a = \boldsymbol{\omega}_a\cdot\hat{\mathbf{k}} = f - \partial u/\partial y + \partial v/\partial x$ is the vertical component of the absolute vorticity $\boldsymbol{\omega}_a$. These types of instability are not the focus of this study and will not be mentioned further.