Ocean Modelling 92 (2015) 69–84

Contents lists available at ScienceDirect

Ocean Modelling

journal homepage: www.elsevier.com/locate/ocemod

The seasonal cycle of submesoscale flows

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ARTICLE INFO

Article history:
Received 9 December 2014
Revised 21 April 2015
Accepted 4 May 2015
Available online 22 May 2015

Keywords:
Submesoscale
Mixed layer fronts
Frontogenesis
Symmetric instability
Baroclinic instability
Seasonal cycle

ABSTRACT

The seasonal cycle of submesoscale flows in the upper ocean is investigated in an idealised model domain analogous to mid-latitude open ocean regions. Submesoscale processes become much stronger as the resolution is increased, though with limited evidence for convergence of the solutions. Frontogenetical processes increase horizontal buoyancy gradients when the mixed layer is shallow in summer, while overturning instabilities weaken the horizontal buoyancy gradients as the mixed layer deepens in winter. The horizontal wavenumber spectral slopes of surface temperature and velocity are steep in summer and then shallow in winter. This is consistent with stronger mixed layer instabilities developing as the mixed layer deepens and energising the submesoscale. The degree of geostrophic balance falls as the resolution is made finer, with evidence for stronger non-linear and high-frequency processes becoming more important as the mixed layer deepens. Ekman buoyancy fluxes can be much stronger than surface cooling and are locally dominant in setting the stratification and the potential vorticity at fronts, particularly in the early winter. Up to 30% of the mixed layer volume in winter has negative potential vorticity and symmetric instability is predicted inside mesoscale eddies as well as in the frontal regions outside of the vortices.

1. Introduction

The upper ocean stratification is an important control on the transfer of momentum and tracers between the atmosphere and ocean interior. The development of upper ocean stratification has historically been viewed as a one-dimensional process driven by surface buoyancy and frictional fluxes, with allowance for shear-driven mixing at the base of the mixed layer. These ideas are encapsulated in a number of one-dimensional parameterisation schemes for the surface boundary layer (e.g. Price et al., 1986; Large et al., 1994). Attention has since focused on the role a number of other processes play in setting upper ocean stratification such as geostrophic adjustment (Dale et al., 2008; Tandon and Garrett, 1994), frontogenesis (Gula et al., 2014; Hoskins and Bretherton, 1972; Lapeyre et al., 2006; Shakespeare and Taylor, 2013), surface waves and Langmuir turbulence (Belcher et al., 2012; Grant and Belcher, 2009; Hamlington et al., 2014; Haney et al., Subm. to JPO; McWilliams and Fox-Kemper, 2013; Sutherland et al., 2014), Ekman buoyancy fluxes (hereafter EBF, Thomas, 2005; Mahadevan, 2006; Thomas and Ferrari, 2008; Thomas et al., 2013), symmetric and inertial instabilities (Bachman and Taylor, 2014; D'Asaro et al., 2011; Haine and Marshall, 1998; Thomas and Taylor, 2010; Thomas et al., 2013; Thomsen et al., 2013), and mixed layer baroclinic instabilities (Bachman and Fox-Kemper, 2013; Boccaletti et al., 2007; Brüggemann and Eden, 2014; Mahadevan et al., 2010; Nurser and Zhang, 2000; Samelson, 1993; Skyllingstad and Samelson, 2012) amongst others. While there is evidence for each of these processes affecting upper ocean stratification, the interactions between them and their relative strength over the seasonal cycle remain major outstanding questions (Belcher et al., 2012; Callies et al., 2015; Capet et al., 2008a; Hamlington et al., 2014; Haney et al., 2012; Lévy et al., 2010; Mensa et al., 2013; Taylor and Ferrari, 2010).

An important point of reference for this work is an insightful series of papers by Capet and co-authors (Capet et al., 2008a; 2008b; 2008c), that examine the transition from mesoscale to submesoscale dynamics in a model domain analogous to the California Current System. An advantage of this approach over a channel model configuration is that the submesoscale processes occur in the context of the strain induced by a larger scale eddy field. This strain may be an important control on the growth rate of instabilities (Bishop, 1993; McWilliams and Molemaker, 2011; Spall, 1997; Thomas, 2012). A comparable experimental methodology is employed in this work whereby simulations are run over a resolution range from mesoscale-resolving to submesoscale-permitting. These simulations depart from previous works in a number of ways. First, a

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http://dx.doi.org/10.1016/j.ocemod.2015.05.002

seasonally varying surface buoyancy forcing is employed and so the mean mixed layer depth varies by an order of magnitude through the year. Second, no temperature-restoring is used and so the model stratification can diverge as the resolution becomes finer. Third, the domain used here is analogous to an open ocean region rather than an eastern boundary current region (Capet et al., 2008a; 2008b; 2008c) or a western boundary current region (Gula et al., 2014; Mensa et al., 2013).

This experiment is carried out in an idealised configuration intended to be analogous to the OSMOSIS (Ocean Surface Mixing - Ocean Submesoscale Interaction Study) observation site in the North Atlantic. The observation site is the Porcupine Abyssal Plain located near (16°W, 49°N) a region where mean flows are weak and mesoscale eddies dominate the kinetic energy budget (Painter et al., 2010). This numerical experiment complements a moored array of instruments, seaglider deployments and two process cruises in the project. Comparisons will be made to these observations as the results are presented, though we note the model has not been 'tuned' to replicate the observations.

This paper is structured as follows. The experimental set-up is given in Section 2. The structure of the buoyancy and velocity fields and the balance relationships that connect them are shown in Section 4. The magnitude of the different submesoscale processes across the seasonal cycle in Section 4. A summary and discussion of the implications for efforts to observe and parameterise submesoscale flows follow in Section 5.

2. Experimental set-up

2.1. Model domain

The simulations are integrated using the MITgcm (Marshall et al., 1997) in a hydrostatic configuration. The model set-up is analogous to the OSMOSIS observation area at the Porcupine Abyssal Plain site. As such, the configuration is that of an open ocean location in the mid-latitudes where the kinetic energy budget is dominated by mesoscale eddies. The domain is doubly-periodic with side-length of 256 km. The bottom boundary is at 3700 m depth and the model domain is spanned with 200 vertical levels. The vertical grid-spacing is reduced near the top and bottom boundaries to 3 m to better resolve the boundary layer processes of interest and increases gradually to a maximum of 32.5 m in the interior.

A series of simulations are carried out with uniform horizontal grid resolutions of 4 km, 2 km, 1 km and 0.5 km. The 4 km run acts as the control for our experiment, though comparisons are also made with observations to ensure the model state is a reasonable representation of the real ocean. The simulations are run on the UK ARCHER supercomputer, a Cray XC30 system. All of the runs are integrated for at least five years with the fifth year used to perform the analysis.

2.2. Numerical configuration

A linear equation of state in temperature is employed with a thermal expansion coefficient $\alpha = 2 \times 10^{-4}\ \mathrm{K}^{-1}$ and so $b = g\alpha(T - T_{ref})$ where b is buoyancy, $g = 9.81\ \mathrm{m\,s^{-2}}$ is gravity, T is temperature and T_{ref} is a reference temperature. Simulations of geostrophic turbulence generate a downscale cascade of enstrophy that must be dissipated to prevent it accumulating at the grid-scale. Enstrophy is dissipated in the momentum equation using adaptive viscous schemes first developed by Smagorinsky (1963), Leith (1996) and Fox-Kemper and Menemenlis (2013). Recent results show that adaptive viscous schemes are necessary to allow submesoscale turbulence to develop (Graham and Ringler, 2013; Ilicak et al., 2012; Ramachandran et al., 2013). Diffusion is applied to horizontal gradients in temperature. For both horizontal diffusion and viscosity, biharmonic operators are chosen over Laplacian operators so that explicit diffusion and viscosity are targeted at the highest wavenumbers (e.g. Griffies and Hallberg, 2000; Graham and Ringler, 2013). At all resolutions the Smagorinsky coefficient is 3, while the Leith and modified Leith coefficients are 1. The biharmonic temperature diffusion coefficient is $4 \times 10^7\ \mathrm{m^4\ s^{-1}}$ at 4 km resolution and reduced by a factor of four for each doubling in resolution. A partial-slip bottom boundary condition is imposed with a quadratic bottom drag (Arbic and Scott, 2008) using a non-dimensional quadratic drag coefficient of 3×10^{-3}.

In addition, vertical mixing of both heat and momentum is carried out with a Laplacian operator with a constant diffusion coefficient of $4 \times 10^{-5}\ \mathrm{m^2\ s^{-1}}$. The mixed layer depth is defined throughout as the first depth where the temperature difference from the surface is greater than 0.1 °C.

The advection of temperature is carried out using the Prather scheme (Prather, 1986). This is an upwind scheme that conserves second-order moments in sub-grid tracer distributions and so helps to preserve the sharp frontal structures of interest. Hill et al. (2012) show that the effective diffusivity of the Prather scheme is similar to the level of diffusion estimated for the real ocean by tracer release studies. The model's default second-order centered advection scheme is employed for momentum.

The timestep is 400 s at 4 km resolution and is then reduced by a factor of two with each doubling in resolution. The model is integrated on an f-plane with a Coriolis frequency $f = 10^{-4}\ \mathrm{s^{-1}}$. Note that no temperature relaxation conditions are employed and so the model solution can evolve freely.

2.3. Boundary layer parameterisation

In the vertical, the model is run with the K-profile parameterisation (*KPP*, Large et al., 1994) for the surface boundary layer. This scheme is in practice a suite of parameterisations that aim to represent a number of mixed layer processes. The *KPP* scheme increases the vertical viscous/diffusive coefficients (hereafter 'diffusive coefficients') based on the surface wind stress. It also increases the diffusive coefficients if there is elevated shear at the base of the mixed layer based on a Richardson number criteria. In the event of destabilising surface buoyancy forcing the *KPP* scheme introduces a vertical non-local transport to capture the effect of vertical convective mixing (Marshall and Schott, 1999). The *KPP* scheme also applies higher diffusive coefficients in the event of negative stratification, even if this is not associated with destabilising surface buoyancy forcing as can occur in the presence of down-front winds. In these cases of static instability the *KPP* scheme applies a high ($5\times10^{-3}\ \mathrm{m^2\ s^{-1}}$) vertical diffusion coefficient rather than instantaneously mixing buoyancy as done by the default MITgcm convective adjustment scheme or the Price et al. (1986) scheme.

2.4. Initial and boundary conditions

The model is initialised at rest with a horizontally uniform temperature profile. The initial vertical temperature profile (Fig. 1, left panel) is derived from an Argo float near the Porcupine Abyssal Plain observation site. This profile was sampled on 23rd March 2012 and is selected as a temperature profile with minimal signs of internal wave heaving or instrument noise.

The model is forced at the surface by a heat flux and wind forcing. The prescribed heat flux is uniform across the domain and averages to zero over each 360-day year (Fig. 1, right panel) with values based on the sum of the net shortwave, longwave, sensible and latent heat fluxes from the monthly climatology of Berry and Kent (2009) for the Porcupine Abyssal Plain observation region. These heat fluxes are applied to the uppermost model level. As such, heating fluxes result in a more rapid restratification than in the real ocean where shortwave radiative fluxes penetrate in an exponentially decaying manner