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Using a resolution function to regulate parameterizations of oceanic mesoscale eddy effects

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ABSTRACT

Mesoscale eddies play a substantial role in the dynamics of the ocean, but the dominant length-scale of these eddies varies greatly with latitude, stratification and ocean depth. Global numerical ocean models with spatial resolutions ranging from 1° down to just a few kilometers include both regions where the dominant eddy scales are well resolved and regions where the model's resolution is too coarse for the eddies to form, and hence eddy effects need to be parameterized. However, common parameterizations of eddy effects via a Laplacian diffusion of the height of isopycnal surfaces (a Gent–McWilliams diffusivity) are much more effective at suppressing resolved eddies than in replicating their effects. A variant of the Phillips model of baroclinic instability illustrates how eddy effects might be represented in ocean models. The ratio of the first baroclinic deformation radius to the horizontal grid spacing indicates where an ocean model could explicitly simulate eddy effects; a function of this ratio can be used to specify where eddy effects are parameterized and where they are explicitly modeled. One viable approach is to abruptly disable all the eddy parameterizations once the deformation radius is adequately resolved; at the discontinuity where the parameterization is disabled, isopycnal heights are locally flattened on the one side while eddies grow rapidly off of the enhanced slopes on the other side, such that the total parameterized and eddy fluxes vary continuously at the discontinuity in the diffusivity. This approach should work well with various specifications for the magnitude of the eddy diffusivities.

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1. Introduction

Mesoscale eddies are ubiquitous in the ocean, and are of leading order importance to the dynamics of major current systems, such as the Antarctic Circumpolar Current (e.g. Hallberg and Gnanadesikan, 2006), the Kuroshio (e.g. Waterman et al., 2011), and the Gulf Stream (e.g. Chassignet and Marshall, 2008). Credible models of the ocean's dynamics need to either explicitly resolve eddies or to parameterize their effects.

The dominant spatial scales of baroclinic ocean mesoscale eddies can be broadly characterized by the first baroclinic deformation radius, which is the distance that a nonrotating first-mode internal gravity wave would propagate in one inertial timescale (e.g. Gill, 1982). With an appropriate regularization at the equator,[1] the first baroclinic deformation radius is given by $L_{Def} = \sqrt{c_g^2/(f^2 + 2\beta c_g)}$, where c_g is the first-mode internal gravity wave speed, f is the Coriolis parameter, and $\beta = \frac{\partial f}{\partial y}$ is its meridional gradient.

In idealized models of baroclinic instability, the upper and lower bounds of unstable wavelengths are proportional to the deformation radius, while the most unstable wavenumber is the inverse of the deformation radius (see, e.g., the textbook by Pedlosky (1987)). The observed dominant eddy length-scales in the ocean vary more slowly with latitude than does the first baroclinic deformation radius (Stammer, 1997), but this may reflect the greater influence of higher baroclinic modes in the tropics and of effectively barotropic eddies in higher latitudes.

Numerical ocean models need to represent the effects of mesoscale eddies, either by explicitly resolving them or via a suitable parameterization, if they are to replicate the dynamical response of the real ocean. As will be illustrated later, the ratio of a model's grid spacing to the deformation radius gives a good indication of whether a model will be locally capable of explicitly resolving eddy effects. However, as both the deformation radius and an ocean model's grid spacing vary in space, one should ask *where*, not *whether*, a global ocean model can explicitly represent eddies. Fig. 1 shows the ocean model horizontal resolution required for the baroclinic deformation radius to be twice the grid spacing, based on a nominally eddy permitting ocean model after one year

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[1] This is just a simple function that goes smoothly between the appropriate equatorial and mid-latitude definitions of the deformation radius without the need for any arbitrary transition latitude (see, e.g. Chelton et al., 1998).

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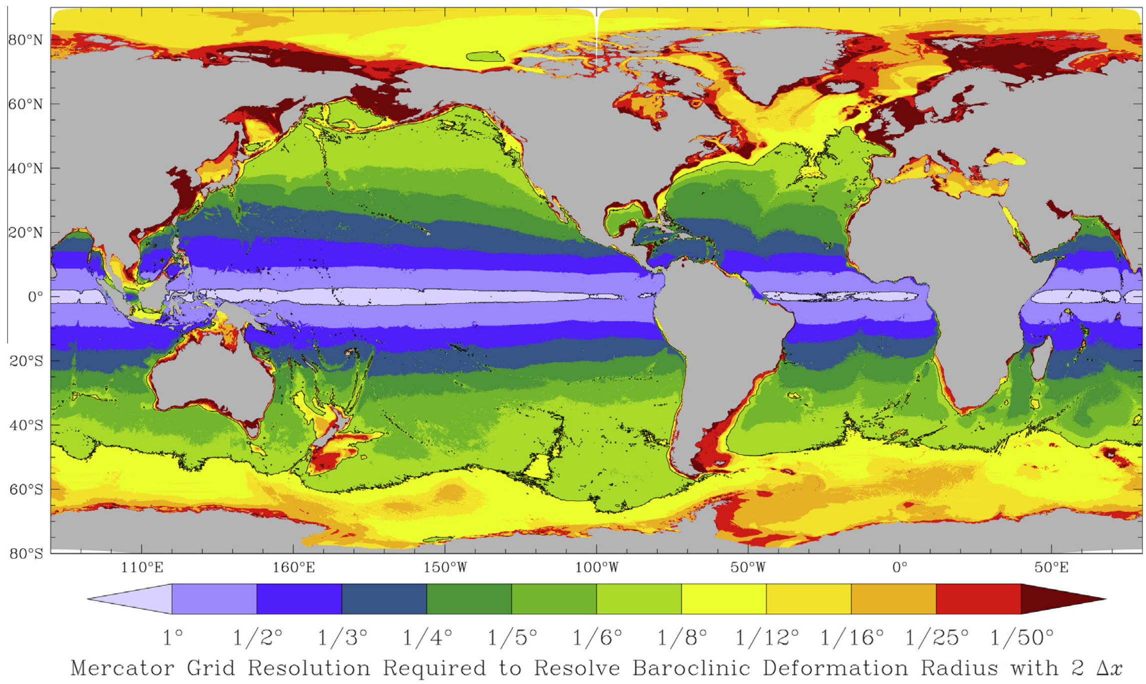

Fig. 1. The horizontal resolution needed to resolve the first baroclinic deformation radius with two grid points, based on a 1/8° model on a Mercator grid (Adcroft et al., 2010) on Jan. 1 after one year of spinup from climatology. (In the deep ocean the seasonal cycle of the deformation radius is weak, but it can be strong on continental shelves.) This model uses a bipolar Arctic cap north of 65°N. The solid line shows the contour where the deformation radius is resolved with two grid points at 1° and 1/8° resolutions.

of spin-up from climatology. At the coarse resolution that is typical of the ocean components of CMIP5 coupled climate models (nominally 1° resolution), an ocean model only resolves the deformation radius in deep water in a narrow band within a few degrees of the equator; any important extratropical eddy effects will need to be parameterized. At a much higher resolution, such as a 1/8° Mercator grid, the deformation radius is resolved in the deep ocean in the tropics and mid-latitudes, but even in this case eddies are not resolved on the continental shelves or in weakly stratified polar latitudes. An unstructured and adaptive grid ocean model could help to address this issue, but such models are not yet in widespread use for global ocean climate modeling, and even then computational speed may dictate the use of models that do not resolve mesoscale eddies everywhere.

In this paper, a series of numerical simulations of a variant of the Phillips (1954) model of baroclinic instability are used to examine the effects of resolution on a numerical model's ability to exhibit the net overturning circulation driven by mesoscale eddies. The effects of a commonly used parameterization of eddy effect, both on the models' explicitly resolved eddies and on the net overturning, are examined. Based on these results, a simple prescription is offered for the typical situation in global ocean models, where eddies are resolved in only part of the domain and in that portion it is desired that the model be allowed to explicitly simulate their effects, but in the remainder of the domain that eddies be entirely parameterized. Specifically, the eddy diffusivities should be multiplied by a "resolution function", ranging from 0 to 1, of the ratio of the baroclinic deformation radius to the model's effective grid spacing, $\widetilde{\Delta} = \sqrt{(\Delta x^2 + \Delta y^2)/2}$. The resolution function that works best for the cases presented here rapidly makes a transition from 1 when this ratio is greater than a value of about 2 (the exact value is not very important and can be chosen to be higher) to 0 for larger values. In the idealized case presented here, this prescription is found to give a reasonable representation of the net eddy-driven overturning over a wide range of resolutions.

2. The test configuration and model

Phillips (1954) analyzed the baroclinic instability that arises in a simple two-layered quasigeostrophic model of a geostrophically sheared flow in a reentrant channel. This problem has the advantage that many of the properties of the eddies, including necessary conditions for the growth of instabilities, the growth rate, energetics and vertical structure of the exponentially growing linear modes can be calculated analytically, as has been documented in many textbooks on geophysical fluid dynamics (e.g. Pedlosky, 1987; Vallis, 2006).

This study examines instabilities of a stacked shallow water variant of the Phillips problem, which is described by the momentum and continuity equations:

$$\frac{\partial \mathbf{u}_n}{\partial t} + \left(f + \hat{\mathbf{k}} \cdot \nabla \times \mathbf{u}_n\right) \times \mathbf{u}_n = -\nabla\left(M_n + \frac{1}{2}\|\mathbf{u}_n\|^2\right) - \nabla \cdot \mathbf{T} - \delta_{n2} c_D \|\mathbf{u}_2\| \mathbf{u}_2, \quad (1)$$

$$\frac{\partial h_n}{\partial t} + \nabla \cdot (h_n \mathbf{u}_n) = (3 - 2n)\left[\gamma\left(\overline{\eta_{3/2}}^x - \eta_{3/2,\mathrm{Ref}}\right) - \nabla \cdot \left(K_h \nabla \eta_{3/2}\right)\right]. \quad (2)$$

Here $\mathbf{u}_n$ is the horizontal velocity in layer n, where $n = 1$ for the top layer and $n = 2$ for the bottom layer. $h_n = \eta_{n-1/2} - \eta_{n+1/2}$ is the thickness of layer n, which is bounded above and below by interfaces at heights $\eta_{n-1/2}$ and $\eta_{n+1/2}$. These equations are solved in a 2000 m deep channel that is 1200 km long and reentrant in the x-direction, and 1600 km wide in the y-direction with vertical walls at the northern and southern boundaries. The Coriolis parameter, f, varies linearly in the y-direction between $6.49 \times 10^{-5}\ \mathrm{s}^{-1}$ and $9.69 \times 10^{-5}\ \mathrm{s}^{-1}$, following the common β-plane approximation. The horizontal stress tensor, $\mathbf{T}$, is parameterized with a shear and resolution dependent Smagorinsky biharmonic viscosity (Griffies and Hallberg, 2000). The Montgomery potentials, $M_n = p/\rho_0 + gz$, in the two layers are given by a vertical integration of the hydrostatic equation, so that