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A multi-model study of the restratification phase in an idealized convection basin

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ABSTRACT

The representation of baroclinic instability in numerical models depends strongly upon the model physics and significant differences may be found depending on the vertical discretization of the governing dynamical equations. This dependency is explored in the context of the restratification of an idealized convective basin with no external forcing. A comparison is made between an isopycnic model including a mixed layer (the Miami Isopycnic Coordinate Ocean Model, MICOM), its adiabatic version (MICOM-ADIAB) in which the mixed layer physics are removed and the convective layer is described by a deep adiabatic layer outcropping at the surface instead of a thick dense mixed layer, and a *z*-coordinate model (OPA model).

In the absence of a buoyancy source at the surface, the mixed layer geometry in MICOM prevents almost any retreat of this layer. As a result, lateral heat exchanges in the upper layers are limited while mass transfers across the outer boundary of the deep convective mixed layer result in an unrealistic outward spreading of this layer. Such a widespread deep mixed layer maintains a low level of baroclinic instability, and therefore limits lateral heat exchanges in the upper layers over most of the model domain. The behavior of the adiabatic isopycnic model and *z*-coordinate model is by far more satisfactory although contrasted features can be observed between the two simulations. In MICOM-ADIAB, the more baroclinic dynamics introduce a stronger contrast between the surface and the dense waters in the eddy kinetic energy and heat flux distributions. Better preservation of the density contrasts around the dense water patch maintains more persistent baroclinic instability, essentially associated with the process of dense water spreading. The OPA simulation shows a faster growth of the eddy kinetic energy in the early stages of the restratification which is attributed to more efficient baroclinic instability and leads to the most rapid buoyancy restoring in the convective area among the three simulations. Dense water spreading and warm surface capping occur on fairly similar time scales in MICOM-ADIAB although the former is more persistent that the latter. In this model, heat is mainly transported by anticyclonic eddies in the dense layer while both cyclonic and anticyclonic eddies are involved in the upper layers. In OPA, heat is mainly brought into the convective zone through the export of cold water trapped in cyclonic eddies with a strong barotropic structure. Probably the most interesting difference between the *z*-coordinate and the adiabatic isopycnic model is found in the temperature distribution ultimately produced by the restratification process. OPA generates a spurious volume of intermediate water which is not seen in MICOM-ADIAB where the volume of the dense water is preserved.

1. Introduction

A large part of the deep water masses of the world ocean are formed in semi-enclosed basins (the Labrador Sea, the Greenland Sea, or the Mediterranean Sea). Observations and numerical simulations conducted in these regions show that the convection can be divided into three phases: a preconditioning phase during which the cyclonic gyre-scale circulation rises isopycnals at the centre of the gyre, bringing the weakly stratified deep water close to the surface, which then appears as a thick homogeneous dense water lens underneath the relatively thin stratified surface layer; a mixing phase initiated by intense surface cooling which erodes the stratified surface water and, for a large enough buoyancy loss, makes it overturn in several plumes and rapidly mix to form a deep homogeneous convective patch; a restratification phase during which a new stratified water column occupying the upper and intermediate layers of the convective patch is established.

Two mechanisms contribute to the restratification, buoyancy added to the surface through heating from the atmosphere or sea ice melt water release, and lateral advection of buoyant stratified water from the periphery of the convective patch. When the mean circulation follows predominantly a circular path around the convective region, lateral exchange must occur primarily via

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mesoscale eddies generated through baroclinic instability of the mean flow (Morawitz et al., 1996). The process is efficient if a buoyant reservoir can be maintained at the periphery of the convective zone through, e.g., a warm boundary current (Lilly et al., 1999). After being restratified, the convective basin is capped by a light surface layer overlying a thick lens of dense homogeneous water, which is a remnant of the previous mixing event. The resulting stratification, especially the characteristics and the thickness of the upper stratified layer, depends upon the efficiency of the restratification processes and ultimately determines the ability of the water column to be destabilized upon entering a new convection cycle.

The dynamics of the convection patch controls the restratification process. The instability builds on the potential energy stored in the front separating the outcropping dense, homogeneous water from the surrounding stratified waters, and, to a lesser extent, on the large scale kinetic energy associated with the rim current in thermal wind balance with this front. Deformation of the mean cyclonic circulation pattern rapidly occurs in which the energy of the large scale flow cascades to smaller scale perturbations. If the radius of the mixed patch is greater than the Rossby deformation radius, the patch is baroclinically unstable and the potential energy of the perturbations is converted into kinetic energy (Marshall and Schott, 1999). Barotropic instability may also occur, in relation to the horizontal shear of the rim current. Several eddies are formed which transport buoyant water from the surroundings to the interior of the convective patch whereas the cold dense water is exported from the convective patch towards the periphery. The role of the eddy fluxes in the restratification has been partially documented by observations in the different convective regions of the world ocean. These suggest that eddy fluxes alone could explain the restratification without the need for an additional buoyancy input from the atmosphere (Send et al., 1995; Lilly et al., 2003). From data collected in the 1990s, Lilly et al. (2003) estimated the eddy contribution to the lateral heat exchange between the boundary and the interior of the Labrador Sea to about 25%.

Analytical considerations together with numerical simulations and laboratory experiments have also demonstrated the central role of geostrophic eddies associated with the baroclinic instability of the rim current in controlling the restratification (Jones and Marshall, 1997). Eddies are believed to constrain the restratification time scale as well as the characteristics of the end products and, provided the restratification operates during the mixing phase, the depth of the convective layer (Visbeck et al., 1996). Smaller scale ageostrophic instabilities, known as mixed layer instabilities, occurring in the less stratified interior region of the mixed patch are also believed to contribute to the restratification, yet their existence is subject to the presence of spatial heterogeneities in the ML density distribution (e.g., Boccaletti et al., 2007). A key issue when addressing the restratification of a convective basin is to understand the link between the eddy heat fluxes and the eddy kinetic energy distribution. Dedicated experiments using more realistic configurations are needed in order to estimate these fluxes and better quantify their impact on the stratification of the basin in relation to the different model physics.

The representation of flow properties and the associated eddy field in numerical models is much dependent upon the model physics and most notably on the parameterization of the subgrid-scale processes (e.g., Willebrand et al., 2001). In particular, major differences are seen when different vertical coordinates are used to discretize the governing dynamical equations. In the ocean interior, far from regions of high mixing rates, transport and mixing preferentially occur along isopycnic surfaces while diapycnal mixing remains fairly low (Griffies et al., 2000a). This high ratio ($\sim 10^8$) of isopycnal mixing to diapycnal mixing is essentially guaranteed in isopycnic models where the two-dimensional transport equation is consistent with the adiabatic framework. As a consequence, such models are well known to better perform in tracking water masses. By contrast, in *z*-coordinate models, the advection schemes only guarantee numerical convergence to approximate adiabaticity and numerical truncation errors and horizontal diffusion introduce spurious diapycnal mixing which unphysically alters the characteristics of the advected water masses (Griffies et al., 2000b). The problem is particularly critical in eddy resolving models where a relatively high horizontal resolution, compared to what would be requested to simulate a realistic eddy field and to dissipate the accumulated variance and enstrophy at the cut-off grid scale, appears to be necessary in order to reduce the level of spurious mixing to small acceptable values (Griffies et al., 2000b).

Stability analyses of a jet like stream have shown that the different representations of advection and mixing between isopycnic and *z*-coordinate models also have important consequences for the characteristics of the hydrodynamic instabilities. In the early stage of the instability of a jet like stream, dissipation is expected to reduce the growth rate of the instability. In *z*-coordinate models, spurious horizontal mixing of density occurring across the jet front tends to reduce the instability growth rate as sharp horizontal density gradients can hardly be sustained (Griffiths et al., 2000). On the other hand, implicit diffusion inherent to the transport scheme used in the continuity equation of the isopycnic models may be very large in weakly unstable flows, thus retarding the instability growth (Griffiths et al., 2000). In the non-linear phase of the instability, spurious diapycnal mixing has been shown to be responsible for retarding the eddy cut-off process in a two-layer *z*-coordinate model, leading to an overshoot effect on the eddy heat flux as the typical size of breaking waves tends to increase with increasing dissipation (Drijfhout, 1992).

In *z*-coordinate models, the effect of dissipation on the instability growth may be counteracted by effects of spurious potential vorticity gradients generated across the Rossby wave front. Such gradients which are the result the non-potential-vorticity-conserving advection scheme tend to artificially enhance the wave growth (Drijfhout, 1992). Inadequate vertical discretization of the density field have the same effect (Ikeda and Wood, 1993) while vertical truncation errors in the advection scheme tend to shorten the wavelength of the fastest growing wave in the linear stage of the instability (Griffiths et al., 2000). These errors may also be responsible for a decrease of the mean kinetic energy and associated baroclinicity, and consequently of the barotropic and baroclinic instabilities (Bleck and Boudra, 1986). One should however remind that, due to the coarse vertical resolution of the models, most of the studies focusing on the impact of the vertical discretization often overestimate this impact.

The relative effects of diffusivity and viscosity on the baroclinic stability of the flow need to be clarified as they largely depend on the model set-up. While both viscosity and diffusivity dampen the wave growth, viscosity appears to i58nfluence the wavelength of the fastest growing wave more than diffusivity in the early stage of the instability (Griffiths et al., 2000). As stated by these authors, this result may be dependent on the particular choice of diffusion parameterization. Additionally, while for small dissipation viscosity largely controls the small scale potential vorticity, for larger dissipation diffusivity is expected to primarily influence the large scale vortex stretching and the conversion of available potential energy (Drijfhout, 1992). The relative impact on the heat transport is however difficult to assess due to the non-linear interactions between viscosity and diffusivity effects. One particular aspect to be clarified is the different effects of viscosity and diffusivity on the production and dissipation of the eddy kinetic energy.

Although isopynic models are expected to perform better in an adiabatic fluid, the isopycnic representation is penalized in