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Impacts of regional mixing on the temperature structure of the equatorial Pacific Ocean. Part 1: Vertically uniform vertical diffusion☆

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ABSTRACT

We investigate the sensitivity of numerical-model solutions to regional changes in vertical diffusion. Specifically, we vary the background diffusion coefficient, κ_b, within spatially distinct subregions of the tropical Pacific, assess the impacts of those changes, and diagnose the processes that account for them.

Solutions respond to a diffusion anomaly, $\delta\kappa_b$, in three ways. Initially, there is a fast response (several months), due to the interaction of rapidly-propagating, barotropic and gravity waves with eddies and other mesoscale features. It is followed by a local response (roughly one year), the initial growth and spatial pattern of which can be explained by one-dimensional (vertical) diffusion. At this stage, temperature and salinity anomalies are generated that are either associated with a change in density ("dynamical" anomalies) or without one ("spiciness" anomalies). In a final adjustment stage, the dynamical and spiciness anomalies spread to remote regions by radiation of Rossby and Kelvin waves and by advection, respectively.

In near-equilibrium solutions, dynamical anomalies are generally much larger in the latitude band of the forcing, but the impact of off-equatorial forcing by $\delta\kappa_b$ on the equatorial temperature structure is still significant. Spiciness anomalies spread equatorward within the pycnocline, where they are carried to the equator as part of the subsurface branch of the Pacific Subtropical Cells, and spiciness also extends to the equator via western-boundary currents. Forcing near and at the equator generates strong dynamical anomalies, and sometimes additional spiciness anomalies, at pycnocline depths. The total response of the equatorial temperature structure to $\delta\kappa_b$ in various regions depends on the strength and spatial pattern of the generation of each signal within the forcing region as well as on the processes of its spreading to the equator.

1. Introduction

Ocean general circulation models (OGCMs) often misrepresent basic features of the density field in the tropical Pacific Ocean, including (i) the location and intensity of the cold tongue in the eastern, equatorial ocean and (ii) the sharpness of the tropical thermocline and near-equatorial fronts. These deficiencies are consequential in that they may lead to errors in simulations of climate variability by coupled general circulation models, for example, contributing to inaccurate representations of near-equatorial currents and the strength and time scale of El Niño–Southern Oscillation (ENSO). A possible cause for these stratification errors is inaccurate parameterizations of mixing processes. The parameterization of subsurface vertical (diapycnal) diffusion is particularly important because it can modify density and pressure, and hence is dynamically active. Furthermore, resolving the small-scale processes responsible for vertical mixing (e.g., Kelvin-Helmholtz instability, internal wave breaking) in OGCMs is impossible in the foreseeable future, and so improving vertical-mixing parameterizations remains a first-order problem.

Parameterizations of subsurface vertical diffusion are commonly represented by a background diffusivity with a coefficient, κ_b, that is constant everywhere or a prescribed function of depth.

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Recently, parameterizations have been tested in which κ_b is allowed to vary spatially, for example, depending on external variables such as mean tidal amplitude, bottom roughness, features of the ocean circulation, or geographical location (e.g., Hasumi and Suginohara, 1999; Jayne and St. Laurent, 2001; Friedrich et al., 2011). The majority of these studies are concerned about effects of vertical mixing in the deep ocean.

Several studies have begun to explore how regional changes in κ_b impact the upper, tropical ocean. It has been recognized that, below the mixed layer and within the region in which κ_b is changed, the response of the stratification (temperature or density) is qualitatively consistent with changes generated locally by the anomalous, vertical diffusive flux (e.g., Richards et al., 2009; Jochum, 2009). Anomalies generated by such local processes are then propagated to other regions by advection, diffusion, or wave radiation. In particular, it has been suggested that off-equatorial effects of diffusion are propagated to the equator by the Pacific Subtropical Cells (STCs; McCreary and Lu, 1994) through advection in the main pycnocline (e.g., Jochum, 2009; Tatebe and Hasumi, 2010; Manucharyan et al., 2011). The equatorial stratification anomalies due to local and remote κ_b changes affect the climate state and variability such as ENSO in atmosphere-ocean coupled models (Meehl et al., 2001; Richards et al., 2009; Jochum, 2009; Manucharyan et al., 2011; Sasaki et al., 2013; Kim et al., 2014).

In this paper, we continue the effort to understand impacts of spatially-varying vertical diffusion in the tropical Pacific. Our goal is to understand the basic processes by which the ocean responds both locally and remotely to changes in κ_b in different regions. For one thing, this knowledge allows the identification of regions where vertical mixing has the greatest impact on important aspects of the ocean state, such as tropical sea surface temperature (SST). For another, it will help in the development of new κ_b parameterizations, by allowing researchers to understand better how the parameterization will impact the ocean state.

We consider κ_b anomalies that are depth independent, the simplest choice when not dealing with particular mixing processes. (We will consider the impact of depth-dependent κ_b anomalies in a companion study; see the discussion at the end of Section 4.) Our approach is to obtain a set of OGCM solutions in which κ_b is increased from a standard value κ_0 to $\kappa_0 + \delta\kappa_b(x, y)$ in spatially distinct subregions of the tropical Pacific, to assess the impact of those changes, and to diagnose the processes that cause them. A particular focus is on how $\delta\kappa_b$ affects the equatorial temperature structure, because the mean climate and its variability are known to be sensitive to that structure. An important aspect of our analysis is that we split temperature anomalies into parts that are either associated with a density change (dynamical anomaly) or without one (spiciness anomaly). As we shall see, these anomalies differ locally from region to region, and they propagate about the basin in very different ways, namely, by radiation of Rossby and Kelvin waves and by advection, respectively.

The paper is organized as follows. Section 2 reports our overall experimental design and describes the various measures that we use to quantify differences between model solutions. Section 3 describes our control run, discusses the processes that adjust solutions to equilibrium in response to forcing by $\delta\kappa_b$, describes the stratification anomalies that develop in several of the regional solutions, and reports the contribution of individual solutions to equatorial SST. Section 4 provides a summary and discussion of results. Appendix A gives precise definitions of the measures of differences, describes how we calculate them, and discusses their properties. Appendix B discusses the properties of regional solutions not reported in Section 3.

2. Experimental design

This section reports our overall approach. We first describe our ocean model and then the suite of solutions that we obtain. We conclude by defining the various measures of solution differences that we use in Section 3.

2.1. Ocean model

We use the Massachusetts Institute of Technology general circulation model (MITgcm; Marshall et al., 1997), which solves the incompressible Navier-Stokes equations on a sphere in a hydrostatic mode with an implicit free surface. Our model set-up is based on Hoteit et al. (2008) and Hoteit et al. (2010) with several modifications. The model domain covers the tropical and subtropical Pacific from 26°S–30°N and 104°E–70°W (see Fig.1), with a constant resolution of 1/3° in both the zonal and meridional directions. The model ocean depth and domain boundaries are defined by the ETOPO2 database (http://www.ngdc.noaa.gov/mgg/global/etopo2.html), the latter defined by the 10-m contour with additional manual editing to remove singular water points. Topography in the Indonesian Seas is also manually edited to allow for reasonable mean transports through narrow channels (e.g., McCreary et al., 2007). The model's vertical resolution ranges from 5 m near the surface to 510 m near the bottom with a total of 51 layers.

Closed, no-slip conditions are specified at land boundaries, and a quadratic form of bottom friction with a drag coefficient of 0.002 is applied. The artificial, northern and southern boundaries, as well as a portion of the western boundary located in the Indian Ocean, are open. Near these boundaries, model variables (temperature, salinity, and horizontal velocity) are relaxed to a monthly climatology determined from the German partner of the consortium for Estimating the Circulation and Climate of the Ocean (GECCO) reanalysis (Köhl et al., 2007; Köhl and Stammer, 2008). Specifically, model variables are relaxed to GECCO values at time scales that vary from 1–20 days within 3° of the boundaries.

Subgrid-scale horizontal mixing is parameterized by biharmonic operators with constant coefficients of 4×10^{11} m^4 s^{-1} for viscosity and 2×10^{11} m^4 s^{-1} for tracer diffusion. The K-profile parameterization of Large et al. (1994) is used for vertical mixing, with background coefficients of 10^{-4} m^2 s^{-1} for viscosity and, for our

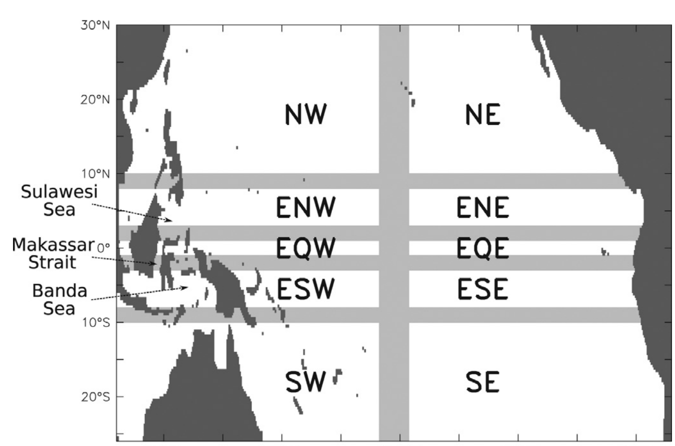

Fig. 1. The model domain showing each geographical region where κ_b is increased from the standard value $\kappa_0 = 0.1 \times 10^{-4}$ m^2 s^{-1} to $\kappa_0 + \Delta\kappa = 0.5 \times 10^{-4}$ m^2 s^{-1}. The gray bands are ramp regions, within which κ_b is ramped sinusoidally from κ_0 to $\kappa_0 + \Delta\kappa$. Precise specifications of these experiments are found in Table 1 and in the text. Two solutions not indicated are Solution CTL (control run) with $\kappa_b = \kappa_0$ everywhere and Solution FB (full basin) in which $\kappa_b = \kappa_0 + \Delta\kappa$ everywhere. Geographical names in the Indonesian Seas are shown for reference in Section 3.3.