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Coastal numerical modelling of tides: Sensitivity to domain size and remotely generated internal tide

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ABSTRACT

The propagation of remotely generated superinertial internal tides constitutes a difficulty for the modelling of regional ocean tidal variability which we illustrate in several ways.

First, the M2 tidal solution inside a control region located along the Southern California Bight coastline is monitored while the extent of the numerical domain is increased (up to 512×512 km). While the amplitude and phase of sea level averaged over the region is quasi-insensitive to domain size, a steady increase of kinetic energy, predominantly baroclinic, is observed with increasing domain size. The increasing flux of energy into the control region suggests that this trend is explained by the growing contribution from remote generation sites of internal tide which can propagate up to the control region.

Increasing viscosities confirms this interpretation by lowering baroclinic energy levels and limiting their rate of increase with domain size. Doubling the grid spacing allows consideration of numerical domains 2 times larger. While the coarse grid has lower energy levels than the finer grid, the rate of energy increase with domain size appears to be slowing for the largest domain of the coarse grid simulations.

Forcing the smallest domain with depth-varying tidal boundary conditions from the simulation in the largest domain produces energy levels inside the control region comparable to those in the control region for the largest domain, thereby confirming the feasibility of a nested approach.

In contrast, simulations forced with a subinertial tidal constituent (K1) show that when the propagation of internal tide is limited, the control region kinetic energy is mostly barotropic and the magnitudes of variations of the kinetic energy with domain size are reduced.

1. Introduction

The oceanic input of tidal energy by astronomical forcing occurs at large spatial scales and the bulk of the response is a barotropic motion which sweeps over the ocean with phase speed exceeding 100 m s^{-1}. In the deep ocean the associated sea level fluctuations and depth-uniform currents are of the order of 1 m and 1 cm s^{-1}, respectively. Tide gauges and satellite altimetry have allowed a detailed mapping of the barotropic response and a better understanding of its dissipation, one third of which is due to the production of baroclinic tidal motion (Egbert and Ray, 2003).

Baroclinic tidal fluctuations are produced when barotropic currents flow across a bathymetric slope and isopycnals are disturbed (Garrett and Kunze, 2007). Guided by maps of barotropic tidal dissipation from satellite altimetry (Egbert and Ray, 2001), observational campaigns near internal tide generation hotspots and numerical simulations have improved our understanding of the generation process over the last decade (Klymak et al., 2006; Legg and Huijts, 2006; Carter et al., 2008). A small fraction of the energy dissipates locally. Most of the energy radiates away as a low mode internal wave (Laurent and Garrett, 2002). For the semidiurnal tide, wavelengths are about 150 km and group speeds are below < 3 m s^{-1} (Alford and Zhao, 2007). The low mode waves can propagate over $O(1000$ km$)$ distances (Dushaw et al., 1995; Ray and Mitchum, 1997; Alford et al., 2007; Zhao and Alford, 2009) and the mechanisms for their ultimate decay are a topic of ongoing debate: bathymetric scattering into higher modes (Bühler and Holmes-Cerfon, 2011), dissipation against coastal boundaries where areas with critical bathymetric slope are abundant (Nash et al., 2004; Martini et al., 2011; Kelly et al., 2012), and nonlinear interaction with the internal wave spectrum (Hazewinkel and Winters, 2011).

Internal tide fluctuations are energetic in the coastal ocean. They are important to marine biology (Lucas et al., 2011), sediment transport (Heathershaw, 1985), lateral heat flux and mass transport (Inall et al., 2001; Shroyer et al., 2010), mixing (Sharples et al., 2007), and acoustic propagation (Duda and Preisig, 1999). The preceding list highlights the need for proper description and prediction of the tidal variability in the coastal domain.

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The long range propagation of tidal fluctuations represents an underestimated challenge for the coastal modeling of tides. A typical study of the three-dimensional tide along the coast uses tidal sea level and current from an assimilation product based on barotropic dynamics (Kurapov et al., 2003; Rosenfeld et al., 2009; Pairaud et al., 2010; Carter, 2010). The effect of remotely generated internal tide is not taken into account, the assumption being that local generation, generally at the shelf break, dominates the variability. This assumption could be justified by the enhanced topographic roughness close to the coasts which could facilitate the reflection and/or scattering and dissipation of remotely generated baroclinic tides before reaching the area of interest. There is evidence that this is not true in general (Martini et al., 2011; Kelly et al., 2012) and it is therefore necessary to verify this assumption, potentially on a case by case basis. One would ideally extend the numerical domain in order to include all possible remote generation sites, but computational resources are ultimately limiting. The present study cannot conclude, for example, on the importance of internal wave sources located more than 1000 km away from our region of interest. Few numerical experiments have investigated the sensitivity of tidal simulations to domain size. Hall and Carter (2011) used simulations on two domains of different sizes (up to 180 $\times$ 180 km) and showed that energy fluxes into the Monterey Canyon are greatest with the larger domain. The present study finds similar results for a different geographical location, extending the results of Hall and Carter (2011). We additionally consider larger domains and investigate how model parameters such as grid spacing and viscosities affect the contribution from remotely generated fluctuations.

We select a control region located in the Southern California Bight offshore San Diego, California and monitor the sensitivity of the local tidal solution to model domain. Most reports on tidal variability in this area have been in depths shallower than 100 m, over the mainland continental shelf (Winant and Bratkovich, 1981; Bratkovich, 1985; Noble et al., 2009; Lucas et al., 2011). Lerczak et al. (2003) reports on observations over the shelf as well as over the shelf break down to 300 m depths. A common feature of the shelf variability is that currents do not tend to follow the spring neap cycle. Lerczak et al. (2003) describes the structure of semidiurnal shelf currents as that of a partially reflected mode 1 wave. Tidal bores have been observed (Pineda, 1994). From a numerical perspective, Buijsman et al. (2012) focused on the Santa Cruz Basin, where the internal tide generation is near resonant and thus is one of the most energetic sites in the Southern California Bight. There have been few other numerical studies of the tidal variability in the area.

The numerical setup is described in Section 2. Section 3 describes the M2 sea level response and its weak sensitivity to domain size. The M2 kinetic energy is next shown to be an increasing function of the domain size (Section 4). This trend is explained by kinetic energy budgets whose inspection in Section 5 reveals the growing amount of baroclinic energy fluxed into the control region when domain size is increased. This is interpreted as the contribution from remotely generated internal fluctuations, which is partially confirmed by the sensitivity of the experiment to viscosity and grid cell size (Section 6). The tidal response of a subinertial constituent (K1) along with its sensitivity to domain size are finally presented in Section 7.

2. Model setup

The control region used by the present study is that of Hoteit et al. (2009), a 30 by 40 km rectangle around Point Loma in San Diego, California (white rectangle in Fig. 1). The numerical calculations have used the MITgcm (Marshall et al., 1997). A set of overlapping numerical domains was chosen such that the eastern edge of the control region is centered on the eastern edge of each domain. The size of the domain doubles from one to the next, from 64 by 64 km (g1, barely bigger than the target area) to 512 by 512 km (g4). The model is run with spherical coordinates and the horizontal grid spacing is approximately 1 km. Vertical grid spacing varies from 1 m close to the surface to 30 m at depth and is the same for all grids. Because the maximum depth increases with numerical domain size, the number of vertical levels varies from 115 (g1) to 200 (g4). Model bathymetry is obtained from the NGDC's 3 arc-second U.S. Coastal Relief Model when available. Elsewhere ETOPO1 (Amante and Eakins, 2009) is linearly interpolated to the model grid. The overlapped grids are aligned so that grid points are collocated horizontally and vertically, and bathymetry is identical in the overlapping portions of the domains. Initial stratification is horizontally uniform, taken from a winter average of CALCOFI station number 28 (http://www.calcofi.org), closest to the target domain (Fig. 1). Below 500 m the temperature and salinity from the 2005 World Ocean Atlas (Locarnini et al., 2006; Antonov et al., 2006) at a nearby deepwater location is used to complete the profile.

Tidal forcing is applied at the boundaries, where the sea level is prescribed. Along-boundary and cross-boundary currents are relaxed on the boundaries to tidally fluctuating values with a 1000 s time scale. This approach differs from the default MITgcm open boundary conditions where the flows normal and tangential to boundaries are prescribed and the sea level adjusts to the flow through boundaries (no boundary values need to be provided for sea level). With this default treatment, the M2 sea level averaged inside the control domain varies with domain size by as much as 8 cm in amplitude and 17° in phase. This is to compare with 2.5 mm and 0.3° when sea level is prescribed along boundaries (see Section 3.2). Note that the choice of default treatment of boundary conditions or prescription of sea level does not affect energy levels by more than 15%. None of the results relative to energy levels presented in this manuscript are qualitatively modified if the default treatment of boundary conditions had been used.

For the largest domain, g4, the model is forced with sea level and barotropic current from the ENPAC tidal database (Spargo et al., 2003). Smaller domains (g1 to g3) are forced by tidal-frequency sea level and currents from a simulation with fixed tracers (g4_noTS see below) on the largest domain. This is done to maintain, at least for the fixed tracer simulations, consistent barotropic dynamics between experiments on different domains. When tracers are freely evolving however, the barotropic dynamics adjusts to some extent from one domain to the next and accommodates for the loss of energy to the internal tide (see Section 5.2).

Temperature and salinity are relaxed to initial profiles within nudging layers along open boundaries. The width of these nudging layers is 5 km for g1 and 10 km for g2 to g3 and the relaxation time scale is 1000 s. The value of the relaxation time scale is set to be smaller than the time for a mode 1 baroclinic wave to cross the width of the nudging layers ($\sim$ 3000 s for a width of 5 km). Sensitivity tests to the relaxation time scale with domain g1 indicate that a value of 1000 s is optimal to minimize baroclinic wave reflections.

The simulations were spun up from rest with the forcing ramped up to full strength over a period of 5 days to reduce transients. The time stepping of sea level is implicit with a 90 s time step. The models are run over a time period of 15 days. Harmonic vertical and horizontal viscosities are constant and with values of 2×10^{-3} and $10\ m^2\ s^{-1}$, respectively. This choice aims at damping grid scale noise and explicitly diagnosing viscous energy loss at the expense of using viscosities much higher than realistic ocean values (Legg and Klymak, 2008; Kelly et al., 2012). Other studies have used more complex turbulence parametrizations (e.g. KPP, Mellor-Yamada 2.0 or 2.5) and/or rather viscous advection schemes. Energy dissipation has to be estimated from the residual