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Active open boundary forcing using dual relaxation time-scales in downscaled ocean models

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ABSTRACT

Regional models actively forced with data from larger scale models at their open boundaries often contain motion at different time-scales (e.g. tidal and low frequency). These motions are not always individually well specified in the forcing data, and one may require a more active boundary forcing while the other exert less influence on the model interior. If a single relaxation time-scale is used to relax toward these data in the boundary equation, then this may be difficult. The method of fractional steps is used to introduce dual relaxation time-scales in an open boundary local flux adjustment scheme. This allows tidal and low frequency oscillations to be relaxed independently, resulting in a better overall solution than if a single relaxation parameter is optimized for tidal (short relaxation) or low frequency (long relaxation) boundary forcing. The dual method is compared to the single relaxation method for an idealized test case where a tidal signal is superimposed on a steady state low frequency solution, and a real application where the low frequency boundary forcing component is derived from a global circulation model for a region extending over the whole Great Barrier Reef, and a tidal signal subsequently superimposed.

1. Introduction

Limited area models nested within larger scale regional models often make use of external data from the regional model in active open boundary conditions, where a radiation condition is relaxed toward the external data with a defined time-scale (Blumberg and Kantha, 1985). Often this relaxation time-scale is adaptive, where long timescales (days to years) are used if the diagnosed wave phase is outgoing near the boundary (which reduces the relaxation term of the radiation condition toward zero and makes the OBC behave more like a pure radiation condition), and short time-scales are used if waves are incoming (Marchesiello et al., 2001). These short timescales must be less than the period of the highest frequency wave present at the boundary; e.g. if a semi-diurnal tide is present in the external data then a timescale of hours may be appropriate.

The relaxation time-scale may be a function of the simulated domain, and previous studies presented in the literature have used a range of different values. Marchesiello et al. (2001) used 1 day and 1 year for incoming and outgoing time-scales respectively when forcing an eastern Pacific shelf model with climatology. Treguier et al. (2001) used 1 day for incoming and 15 and 1500 days for outgoing waves in a model of the Atlantic forced with climatology. Barnier et al. (1998) used 15 days and 5 years for incoming and outgoing waves respectively for a South Atlantic model, again forced with climatology. Blumberg and Kantha (1985) used the time it takes a transient to traverse the shelf as the relaxation time-scale in shelf models (they found 4 h optimum for their application). Gan and Allen (2005) split the open boundary solution into forced (derived from a 2D sub-model) and interior components, diagnosed the phase speed from the interior solution and radiated the interior solution for outgoing phases and imposed the forced solution with timescale of 0.5 days for incoming signals. These relaxation time-scales are constant in space and time; Chen et al. (2013) developed an outgoing timescale diagnosed from the model state that is variable in space and time.

Although adaptive approaches are designed to retain a passive character when phase speed is diagnosed as outgoing, in reality this is not always achieved; phase speeds can resemble random noise (Treguier et al., 2001) and radiation can be ineffective (Herzfeld and Andrewartha, 2012). Additionally, when a tide is imposed together with low frequency oscillations it becomes difficult to apply incoming relaxation timescales that accurately capture the amplitude and phase of the tide whilst allowing the open boundary to retain some degree of passiveness and be transmissive to outgoing transients. Timescales that are too short will

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generate open boundary over-specification and an overly reflective boundary (the boundary essentially becomes clamped), whereas too long time-scales results in boundary under-specification where the tidal signal is not reproduced. Herzfeld et al. (2011) used 30 min incoming and 20 days outgoing for a tidally resolved model of Spencer Gulf, South Australia, and showed when the incoming time-scale was increased to 4 h significant phase shifts were introduced in the tidal signal. In the Great Barrier Reef (GBR) lagoon off north-eastern Australia, at certain times of the year the low frequency response is locally forced by wind, but large tides of amplitude >4 m are also present. Development of an accurate open boundary condition to capture both these processes in the GBR is a motivating factor for this study.

In this study we seek to use relaxation techniques analogous to the approach of spectral nudging, where in this approach only specified frequencies of a prognostic field are nudged toward a supplied climatology to allow simulated eddy fields to evolve without drifting from the climatology (Stacey et al., 2006). In this case we use dual relaxation time-scales, where a short timescale is applied to the high frequency sea-level component (tides) and a longer relaxation time-scale for the low frequency component. We cast this dual relaxation in the open boundary condition of Herzfeld and Andrewartha (2012). This condition is based on a Dirichlet condition rather than radiation, using a local flux adjustment to maintain volume conservation and to introduce the tidal forcing. We cast the local flux adjustment into a form that accepts dual fluxes, which we then split using the method of fractional steps and subsequently apply different time-scales to each component of the split normal boundary fluxes. The method is described in detail in Section 2. We then apply the method to a simple test case in Section 3 to demonstrate its effectiveness, and further to a real domain in Section 4. Finally we draw some conclusions in Section 5.

2. The dual relaxation method

The open boundary condition of Herzfeld and Andrewartha (2012) is a Dirichlet based condition where 3D velocities from an external model are directly applied to the open boundaries of the downscaled model. Barotropic velocity open boundary conditions are the depth average of these 3D velocities. The elevation is left unconstrained in this OBC scheme in the sense that no open boundary equation is applied, and elevation in the boundary cell is solved by the continuity equation identically to the rest of the domain. This allows elevation to perfectly respond to outward propagating signals; it is this aspect of the open boundary condition that reduces the impact of over-specification error. However, since volume fluxes through boundary sections invariably differ between external and downscaled models (due to bathymetry differences and interpolation error), the downscaled model is often prone to basin filling or emptying over time. To overcome this, a local flux adjustment is applied, where normal depth averaged velocities are adjusted every time-step so that the volume divergence in the boundary cell achieves some 'target' elevation (supplied from the external model, and optionally augmented with the tidal signal). The flux adjustment scheme can therefore influence elevation in the boundary cell by adjustment of the depth averaged normal boundary velocity. The normal depth averaged velocity is relaxed toward this adjusted velocity with a pre-defined time-scale. If this relaxation timescale is equal to the 2D time-step, then the target elevation is achieved exactly every time-step and the boundary behaves as a (reflective) clamped condition. In practice, the relaxation timescale is usually provided by the user, noting that using a timescale close to the 2D time-step results in elevation rapidly converging on the supplied target elevation, and a timescale much longer than the 2D time-step results in little influence of the external sea level data. Herzfeld and Andrewartha (2012) noted that if a timescale of $\tau_f = h_1/\sqrt{gD_B}$ is used (h_1 = grid spacing, D_B = depth, g = acceleration due to gravity), then the local flux adjustment takes the form of the Flather radiation condition (Flather, 1976), and this time-scale can be considered a 'default' timescale that optimizes volume conservation and area averaged kinetic energy in the domain (Herzfeld and Andrewartha, 2012). However, this time-scale is too long to accurately reproduce the tidal amplitude and phase characteristics in many domains, and we seek an alternative timescale that optimizes actively applied tidal signals but is passively responsive to low frequency signals generated locally.

Assume that the normal barotropic velocity through the open boundary face is equal to the sum of a high and low frequency component, $V_B^t = V_{BT} + V_{BL}$. The continuity equation can be written in a form where a flux through one cell face is similarly split into contributions from high and low frequency depth averaged velocities, e.g. for a western boundary;

$$\frac{\eta^{t+\Delta t} - \eta^t}{\Delta t_{2D}} = -\nabla DU = -\left[F_{i+1} - (F_T + F_L) + F_{j+1} - F_j\right]/h_1h_2 = \frac{F_T + F_L}{A} - \frac{\Delta F}{A} \quad (2.1)$$

where $U = \frac{1}{\eta - H}\int_H^\eta u dz$ is the depth average of velocity u ($H < 0$ is the depth), $\eta^{t+\Delta t}$ and η^t are the elevations at the forward and current time-step respectively (using Euler forward time discretization), Δ^t_{2D} is the 2D time-step, $D = \eta^t - H$ is total depth, h_1 and h_2 are the grid spacing in the axis directions e_1 and e_2 respectively, F_T is the tidal volume flux through cell face i, F_L is the low frequency volume flux through cell face i, F_{i+1}, F_j and F_{j+1} are volume fluxes through cell faces $i+1$, j and $j+1$ respectively, $\Delta F = F_{i+1} + F_{j+1} - F_j$ and $A = h_1h_2$.

Eq. (2.1) is then set up in two fractional steps (Kowalik and Murty, 1993, p. 64);

$$\frac{1}{2}\frac{\eta' - \eta^t}{\Delta t_{2D}/2} = \frac{\eta' - \eta^t}{\Delta t_{2D}} = \frac{F_T}{A} - \frac{\Delta F}{2A} \quad (2.2)$$

$$\frac{1}{2}\frac{\eta^{t+\Delta t} - \eta'}{\Delta t_{2D}/2} = \frac{\eta^{t+\Delta t} - \eta'}{\Delta t_{2D}} = \frac{F_L}{A} - \frac{\Delta F}{2A} \quad (2.3)$$

where η' is an intermediate elevation. Then Eq. (2.1) = (2.2) + (2.3). Assume that target tidal and low frequency elevations are supplied, η_T and η_L respectively. Following Appendix A in Herzfeld and Andrewartha (2012), first set the flux required to achieve the elevation due to the tide (η_T) using Eq. (2.2);

$$F_T = \frac{A}{\Delta t_{2D}}(\eta_T - \eta^t) + \frac{\Delta F}{2} \quad (2.4)$$

Given the depth at the boundary at time t (D_B) and cell width in e_2 direction (h_2), the depth averaged velocity required to achieve this flux is;

$$V_T = \frac{F_T}{h_2 D_B} \quad (2.5)$$

The velocity on the boundary is then relaxed towards V_T using a time-scale of τ_T according to (Eq. (A.3) in Herzfeld and Andrewartha, 2012):

$$V'_{BT} = V_B^t - \frac{\Delta t_{2D}}{\tau_T}(V_B^t - V_T) \quad (2.6)$$

Here the prime (′) implies an approximation to the true tidal velocity, V_{BT}. The flux through the boundary cell to account for tidal motion, $F_{BT} = V'_{BT}h_2D_B$, is then computed, followed by a new elevation in the boundary cell using Eq. (2.2);