ROAM : Task 5 Model Description and Assessment

M. Herzfeld
CSIRO Marine and Atmospheric Research

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Executive Summary

A relocatable model (ROAM: Relocatable Ocean and Atmosphere Model) was developed to simulate the hydrodynamics of limited area domains in the Australisian region. This model is to run operationally, forecasting up to 10 days. ROAM is forced with initial and open boundary conditions generated offline by a global ocean model (OFAM), and by atmospheric surface fluxes generated by a relocatable atmospheric model (RAMS).

ROAM requires minimal user input, consisting only of the geographic bounds and resolution of a region, and the simulation start / stop time. Based on this information, the parameterisations required for the simulation to proceed are computed by the model. ROAM is required to be unconditionally stable, i.e. solutions are required to be generated by the model on a first-time, every-time basis. The parametrisation used for the model therefore favoured robustness in favour of accuracy. Even so, the model was found to become unstable in about 25% of the 170 simulations performed on 95 different regions encompassing a range of resolutions, geographies, forcing regimes and dominant physics. Most of the regions where instability was observed featured strong non-linear flow, such as in boundary currents or the Indonesian throughflow. A strategy of targeted constraint was employed to ensure ROAM achieved unconditional stability. This involved establishing thresholds on velocity and elevation, and performing targeted filtering or relaxation when those thresholds were locally exceeded. Statistics relating to the 170 simulations formed the basis for determining the velocity and elevation thresholds. The potential exists to further fine tune these thresholds as more experience is gained with ROAM.

In the majority of simulations performed, the model solutions were free from obvious numerical error, and temperature and salinity (T/S) solutions exhibited similar general features to the global model. ROAM T/S solutions generally displayed more structure than OFAM T/S, especially as resolution increased. In some cases error was observed in the form of grid-point noise due to insufficient horizontal viscosity, open boundary re-circulations, unrealistically large velocities, or T/S smearing due to the open boundary treatment.

The robust parameterisation in conjunction with constraining methods has proven to achieve unconditional stability. However, the solutions cannot be considered unconditionally accurate. Occasionally, the model must adaptively shift to the robust end of the robustness - accuracy spectrum to preserve stability, with associated consequences for accuracy.
1 Introduction

The recent development of numerical ocean models has seen a large number of different codes applied to a variety of domains encompassing a spectrum of spatial and temporal scales. At shelf to estuarine scales, these domains may be represented by limited-area models which employ open boundaries to propagate basin scale phenomena into the domain using nesting techniques (e.g. Penven et al, 2006, Onken et al, 2005). Although increased computing power has seen the resolution of general circulations models increase, the global models cannot yet provide solutions at the resolution that is often required to study coastal phenomena. The limited-area approach involves location of the model to the domain of interest, where appropriate grid specification, bathymetry, open-boundary configuration, vertical discretisation and parameterisation of forcing and mixing processes must be specified. These specifications are usually different for different applications, and are often challenging to establish.

Once the model domain is set up, the time steps, diffusion coefficients, roughness parameters and open-boundary configurations are selected to ensure that the model is stable while executing. Since the open-boundary problem is ill-posed, there are no easy rules for choosing the conditions on each new application with new bathymetry, geography, dominant physics or forcing mechanisms. The choice of advection scheme may also impact stability, especially if the scheme used is not positive definite or monotonic.

CSIRO Marine and Atmospheric Research, in collaboration with the Royal Australian Navy (RAN) and Bureau of Meteorology (BOM) have undertaken to develop a relocatable model nested within a global ocean circulation model (Griffies et al., 2005), and forced with surface fluxes from a relocatable atmosopheric model (RAMS, see Abbs, 2006). The relocatable model (ROAM: Relocatable Ocean and Atmosphere Model) is to run operationally, forecasting out to 10 days. Inputs to the model are only the latitude and longitude bounds of the domain of interest, start/stop times of the simulation, and the horizontal resolution. The operational implementation necessitates that solutions must be generated by the model on a first-time, every-time basis. That is, an unconditionally stable simulation must be achieved for every region simulated. This paper describes the parameterisations and techniques that were employed to achieve unconditional stability for the oceanic model included in ROAM. Section 2 describes this model and section 3 details the methods used to derive parameterisations. Generally, model robustness requires a diffusive model, often run with very small time-steps. Small time-steps can result in unacceptably long model run-times. Large diffusion invariably leads to a loss of accuracy in solutions, since density or velocity structure tends to be smeared out. Therefore, the model may be considered to be positioned at some point on a robustness/accuracy spectrum. Obviously the preferred outcome is to maximise accuracy without compromising stability. However, unconditional stability proves to be an unrealistic expectation for a fixed model parameterisation, since there appears always to be a combination of forcing, dominant physics, model computational physics and parameterisations that may render the model unstable. Unconditional stability was achieved through a strategy of containment rather than robust parameterisation. Section 4 describes the constraining methods employed, followed by a discussion of ROAM applications in section 5.
2 The Model

The oceanic component of ROAM is a general purpose model applicable to scales ranging from estuaries to regional ocean domains, and has been successfully applied to a variety of applications encompassing these scales (e.g. Walker, 1997, Parslow et al. 2001, Herzfeld et al. 2003). It is a three-dimensional finite-difference hydrodynamic model, based on the primitive equations. Outputs from the model include three-dimensional distributions of velocity, temperature, salinity, density, passive tracers, mixing coefficients and sea-level. Inputs required by the model include forcing due to wind, atmospheric pressure gradients, surface heat and water fluxes and open-boundary conditions (e.g. tides). ROAM is based on the equations of momentum, continuity and conservation of heat and salt, employing the hydrostatic and Boussinesq assumptions. The equations are discretized on a finite-difference stencil corresponding to the Arakawa C grid.

The model uses a curvilinear orthogonal grid in the horizontal and a choice of fixed ‘z’ coordinates or terrain following σ coordinates in the vertical. The ‘z’ vertical system allows for wetting and drying of surface cells, useful for modelling regions such as tidal flats where large areas are periodically dry. The bottom topography is represented using partial cells. The model has a free surface and uses mode-splitting to separate the two-dimensional (2D) mode from the three-dimensional (3D) mode. Mode-splitting allows fast moving gravity waves to be solved independently from the slower moving internal waves, allowing the 2D and 3D modes to operate on different time-steps, for computational efficiency. The model uses explicit time-stepping throughout, except for the vertical diffusion scheme which is implicit. A Laplacian diffusion scheme is employed in the horizontal on geopotential surfaces. Smagorinsky mixing coefficients may be utilized in the horizontal.

The ocean model can invoke several turbulence closure schemes, including k-ε, Mellor-Yamada 2.0 and Csanady-type parameterisations. A variety of advection schemes may be used on tracers, and 1st or 2nd order can be used for momentum. This study used the advection scheme of Van Leer (1979). The model also contains a suite of radiation, extrapolation, sponge and direct data forcing open-boundary conditions. In its general use, the model is capable of performing particle tracking and may be directly coupled to ecological and sediment transport models.

The equations of motion are based upon those described in Blumberg and Herring (1987). The equations used in the ‘z’ model are presented here; the σ model follows the formulation of Blumberg and Herring (1987) and is not repeated here. If the three-dimensional vector of velocity has the components $u_1$ in the $\xi_1$ direction, $u_2$ in the $\xi_2$ direction and $w$ in the $z$ direction, then:

\[
\begin{align*}
  u_1 &= h_1 \frac{d\xi_1}{dt} \\
  u_2 &= h_2 \frac{d\xi_2}{dt} \\
  w &= \frac{dz}{dt}
\end{align*}
\]

where $h_1$ and $h_2$ are the metrics in the $\xi_1$ and $\xi_2$ directions respectively. The continuity equation is:
Momentum in the $\xi_1$ direction is given by:

$$\frac{\partial u_1}{\partial t} + \frac{1}{h_1 h_2} \left[ \frac{\partial (u_1^2 h_1 h_2)}{\partial \xi_1} + \frac{\partial (u_1 u_2 h_2)}{\partial \xi_2} \right] + \frac{\partial (wu_1)}{\partial z} = - \frac{1}{h_1 \rho} \frac{\partial P}{\partial \xi_1} + f u_2 + \frac{u_1^2}{h_1^2} \frac{\partial h_1}{\partial \xi_1} + \frac{u_2^2}{h_2} \frac{\partial h_2}{\partial \xi_2} + \psi_1 + \frac{\partial}{\partial z} \left[ V_z \frac{\partial u_1}{\partial z} \right]$$

where $t$ is the time, $P$ is pressure, $\rho$ is the density, $f$ is the Coriolis parameter, $\psi_1$ is the horizontal $\xi_1$ momentum diffusion term and $V_z$ is the vertical viscosity coefficient. Similarly, the momentum in the $\xi_2$ direction is:

$$\frac{\partial u_2}{\partial t} + \frac{1}{h_1 h_2} \left[ \frac{\partial (u_1^2 h_1 h_2)}{\partial \xi_1} + \frac{\partial (u_1 u_2 h_2)}{\partial \xi_2} \right] + \frac{\partial (wu_2)}{\partial z} = - \frac{1}{h_1 \rho} \frac{\partial P}{\partial \xi_2} - f u_1 + \frac{u_1^2}{h_1 h_2} \frac{\partial h_1}{\partial \xi_1} + \frac{u_2^2}{h_2^2} \frac{\partial h_2}{\partial \xi_2} + \psi_2 + \frac{\partial}{\partial z} \left[ V_z \frac{\partial u_2}{\partial z} \right]$$

where $\psi_2$ is the horizontal $\xi_2$ momentum diffusion term (see Blumberg and Herring (1987)).

The conservation equation for any tracer $T$, is given by:

$$\frac{\partial T}{\partial t} + \frac{1}{h_1 h_2} \left[ \frac{\partial (u_1 T h_2)}{\partial \xi_1} + \frac{\partial (u_2 T h_1)}{\partial \xi_2} \right] + \frac{\partial (wT)}{\partial z} = \frac{1}{h_1 h_2} \left[ \frac{\partial}{\partial \xi_1} \left( h_2 V_h \frac{\partial h_2}{\partial z} + \frac{h_1}{h_2} V_h \frac{\partial T}{\partial z} \right) + \frac{\partial}{\partial \xi_2} \left( h_1 V_h \frac{\partial h_1}{\partial z} + h_2 V_h \frac{\partial T}{\partial z} \right) \right] + \frac{\partial}{\partial z} \left[ K_z \frac{\partial T}{\partial z} \right]$$

where $V_h$ is the horizontal diffusion coefficient and $K_z$ is the vertical diffusivity.

Equation 2.7 is valid for temperature, salinity and any contaminants included. The vertical velocity is obtained from Eq. 2.4. The boundary conditions at the surface are:

$$\rho V_z \frac{\partial}{\partial z} (u_1, u_2) = (\tau_{sx}, \tau_{sy})$$

$$\rho K_z \frac{\partial}{\partial z} (T, S) = (H_T, H_S)$$

where $T$ is the temperature, $S$ is the salinity, and $\tau_{sx}$ is the surface wind stress in the $\xi_1$ direction, $\tau_{sy}$ is the surface wind stress in the $\xi_2$ direction, $H_T$ is the surface heat flux and $H_S$ is the surface salt flux. The surface boundary condition for vertical velocity is:
\[
W_{\text{top}} = \frac{\partial \eta}{\partial t} + \frac{u_{1\text{top}}}{h_1} \frac{\partial \eta}{\partial \xi_1} + \frac{u_{2\text{top}}}{h_2} \frac{\partial \eta}{\partial \xi_2}
\]

where \( \eta \) is the surface elevation, \( u_{1\text{top}} \) is the surface \( \xi_1 \) velocity and \( u_{2\text{top}} \) is the surface \( \xi_2 \) velocity.

The boundary conditions at the bottom are:

\[
\rho V_z \frac{\partial}{\partial z} (u_1, u_2) = (\tau_{b\xi}, \tau_{b\eta})
\]

\[
\rho K_z \frac{\partial}{\partial z} (T, S) = 0
\]

\[
W_{\text{bot}} = \frac{u_{1\text{bot}}}{h_1} \frac{\partial H}{\partial \xi_1} - \frac{u_{2\text{bot}}}{h_2} \frac{\partial H}{\partial \xi_2}
\]

where \( \tau_{b\xi} \) is the bottom stress in the \( \xi_1 \) direction, \( \tau_{b\eta} \) is the bottom stress in the \( \xi_2 \) direction, \( u_{1\text{bot}} \) is the bottom \( \xi_1 \) velocity and \( u_{2\text{bot}} \) is the bottom \( \xi_2 \) velocity.

The bottom stress is given by the quadratic friction law:

\[
\tau_{b\xi} = \rho C_D (u_1^2 + u_2^2)^{1/2} u_1
\]

\[
\tau_{b\eta} = \rho C_D (u_1^2 + u_2^2)^{1/2} u_2
\]

where the drag coefficient is given by:

\[
C_D = \max \left( \frac{1}{\kappa} \ln \left( \frac{z + z_0}{z_0} \right), C_{D\min} \right)
\]

Here \( z \) denotes the distance above the sea floor. Numerically this is implemented as the first grid point above the bottom. \( C_{D\min} \) is a minimum drag coefficient (typically between 0.002 and 0.003) used when the first grid point is a long way from the bottom (i.e. for large \( z \)), \( z_0 \) is the bottom roughness and \( \kappa = 0.4 \) is Von Karman’s constant.

Vertical integration of the continuity and momentum equations leads to equations for the 2D mode, vis;
\[
\frac{\partial \eta}{\partial t} + \frac{1}{h_hh_2} \left[ \frac{\partial DU_1}{\partial \xi_1} h_hh_2 + \frac{\partial DU_2}{\partial \xi_2} h_2 \right] = 0
\]

2.16

where \( U_1 \) and \( U_2 \) are the vertically integrated \( u_1 \) and \( u_2 \) velocities respectively and \( D = H + \eta \) is the total depth where \( H \) is the bottom depth. The vertical integrals of eqns. 2.5 and 2.6 give:

\[
\frac{\partial U_1}{\partial t} + \frac{1}{Dh_1^2h_2} \left[ \frac{\partial DU_1^2}{\partial \xi_1} h_hh_2 + \frac{\partial DU_1}{\partial \xi_2} h_2^2 \right] = fU_2 - \frac{1}{h_1\rho_{av}} \left[ \frac{\partial p_a}{\partial \xi_1} + g\rho_{top} \frac{\partial \eta}{\partial \xi_1} \right] - \frac{g}{D} \int_0^H h_1 \rho \frac{\partial \xi_1}{\partial z} dz
\]

2.17

\[
\frac{\partial U_2}{\partial t} + \frac{1}{Dh_2^2h_1} \left[ \frac{\partial DU_2}{\partial \xi_1} h_hh_2 + \frac{\partial DU_2^2}{\partial \xi_2} h_1^2 \right] = fU_1 - \frac{1}{h_2\rho_{av}} \left[ \frac{\partial p_a}{\partial \xi_2} + g\rho_{top} \frac{\partial \eta}{\partial \xi_2} \right] - \frac{g}{D} \int_0^H h_2 \rho \frac{\partial \xi_2}{\partial z} dz
\]

2.18

where

\[
\begin{align*}
        u'_1 &= u_1 - U_1 \\
        u'_2 &= u_2 - U_2
\end{align*}
\]

2.19
3 Parameterisations

3.1 Open boundary conditions

Normal and tangential velocity components, sea-level, and temperature and salinity must be prescribed at the open boundaries of the model. The open-boundary problem has received much attention in the literature (e.g. Chapman, 1985, Roed and Cooper, 1987, Tang and Grimshaw, 1996, Palma and Matano 1998, 2001). The problem is mathematically ill-posed, and no open-boundary condition (OBC) can be considered perfect. Specification of a robust boundary parameterisation is a challenge in many limited-area ocean-model implementations.

The stencil used for open-boundary conditions is illustrated in Figure 3.1. This configuration is similar to that used by Stephens (1990), where the normal component of velocity is confined to the non-linear terms and not allowed to propagate into the interior via the Coriolis term. The normal velocity open-boundary condition is the zero-flux, or no-gradient, condition. This condition has been shown to maintain stability where radiation conditions do not (Palma and Matano, 1998, 2001). Tangential velocity was clamped to zero. Although this OBC is completely reflective (Chapman, 1985), it was found to be more robust than alternative radiation conditions.

A one-way nesting strategy (e.g. Spall and Robinson, 1989) was used to propagate the large-scale circulation into the regional domain. This technique is compatible with operational ocean forecasting (Barth et al, 2005). In this case, the sea-level and baroclinicity were prescribed at the open boundaries. This is an ‘active’ condition for elevation, which may lead to model instability due to reflection of energy at the boundary, or inaccuracies due to over-specification (Marchesiello et al, 2001). There are essentially two strategies for dealing with outgoing perturbations (transient responses) generated inside the computational domain. Radiation conditions make the boundary effectively transparent to the transients, allowing them to propagate out unhindered (e.g. Blumberg and Kantha, 1985). Alternatively, the transients may be damped before they reach the boundary by using sponge layers adjacent to the boundary, in which diffusion is increased (e.g. Israeli and Orszag, 1981). Radiation conditions may be made active by allowing relaxation towards external data. Adaptive relaxation may be applied for inward and outward propagation (Marchesiello et al 2001). However, the response of the model interior is sensitive to the choice of radiation condition used and the relaxation timescale. As noted by Palma and Matano (1998), due to the ill-posed nature of the open-boundary problem, the numerical behaviour of the OBC may vary with the specific circumstances of the simulation, the model characteristics or its numerical
implementation. This makes unconditional stability elusive, since consistent OBC behaviour is not guaranteed in response to subtle changes in aspects of the problem.

A more robust approach is the use of sponges on the boundary, rather than radiation conditions. The sponge took the form of increased horizontal viscosity over a region eight cells wide adjacent to the boundary. Sponges were blended laterally to remove large discontinuities with interior viscosities. Maximum viscosity was derived from the stability constraint for diffusion with a ‘safety factor’ of 0.9 applied, i.e. maximum viscosity took the form:

$$A_{\text{max}} = \frac{0.9}{4\Delta t} \left[ \frac{1}{\Delta x^2} + \frac{1}{\Delta y^2} \right]^{-1}$$

This maximum viscosity was interpolated to the interior values over the width of the sponge using the hyperbolic tangent function:

$$A_{\text{HS}} = A_H + (A_{\text{max}} - A_H)(1 - \tanh(\alpha(x - 1)))$$

where $x$ is the number of cells from the boundary and $\alpha = 0.3$. Sponges were applied separately for $u_1$ and $u_2$ velocities normal to the open boundaries.

The low-frequency sea-level response is provided by the global model at daily intervals. Superimposed on this is the higher frequency sea-level change due to the tide. The global tidal model of Eanes and Bettadpur (1995), using the methodology of Cartwright and Ray (1990), is used to generate amplitudes and phases of 14 tidal constituents at every open-boundary node. These are then used to reconstruct the tide at every time-step. The prescription of elevation and baroclinicity at the boundary is essentially only forcing the model with the pressure term from the equations of motion, so that solutions only retain accuracy if flow is close to geostrophic near the boundary. An alternative approach would directly prescribe velocity from the global model on the boundary. However, since the global model does not include the tide, this is not possible. A global model incorporating the tide would remain impractical, since OBC’s would not be available at a fine enough temporal resolution due to file size and storage.

Temperature and salinity used an upstream advection OBC (Blumberg and Mellor, 1986). This is essentially the one-dimensional advection equation solved using the first-order upwind method. For example, in the $e_1$ direction:

$$T^{n+1} = T^n + \frac{\Delta t}{h_i} \left[ 0.5(u_i + |u_i|)(T_i - T_{i-1}) + 0.5(u_i - |u_i|)(T_B - T_i) \right]$$

When flow is into the domain, external data, $T_B$, are required. In this case, $T_B$ is derived from the global model at daily intervals and interpolated in time.

### 3.2 Bathymetry

A geographic rectangular grid was applied to all domains. This is essentially an orthogonal grid on a spheroid, where grid spacing is defined as increments of latitude and longitude, resulting in unequal grid spacings on the sphere. The grid may be rotated from a north-south
orientation. Bathymetry is interpolated onto the grid using the AGSO (2002) bathymetric
database (Petkovic and Buchanan, 2002) at 1km resolution around the Australasian region,
and the DBDB2 v2.2 (Digital Bathymetry Data Base, 2 minute resolution, US Naval Research
Laboratory, Ocean Dynamics and Prediction Branch, http://www7320.nrlssc.navy.mil/) bathymetric
product for other regions. The bathymetry was globally smoothed using a 9-point
convolution filter and a minimum depth of 4 m was imposed. The minimum depth ensures
that cells always remain wet, even in the north-west regions of Australia where tidal range
can be 3m or more. Finally the bathymetry was smoothed locally if the bathymetric gradient
exceeded 0.05. A gradient of 0.1 is considered an upper limit for values found in the ocean,
(Mellor and Blumberg, 1985). Removal of extreme bathymetric gradients assists model
stability.

### 3.3 Layer structure

The vertical discretisation was derived from a sigma profile applied to the deepest water
column, with exponential spacing from the surface down until the grid spacing becomes
greater than a threshold ($\Delta z_{lim}$), after which linear spacing is applied to the bottom ($D_{max}$). The
thresholds are listed in Table 3.1.

<table>
<thead>
<tr>
<th>$\Delta z_{lim}$</th>
<th>$D_{max}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>$D_{max} \leq 50$</td>
</tr>
<tr>
<td>50</td>
<td>$50 &lt; D_{max} \leq 200$</td>
</tr>
<tr>
<td>100</td>
<td>$200 &lt; D_{max} \leq 1000$</td>
</tr>
<tr>
<td>150</td>
<td>$1000 &lt; D_{max} \leq 2500$</td>
</tr>
<tr>
<td>200</td>
<td>$D_{max} &gt; 2500$</td>
</tr>
</tbody>
</table>

*Table 3.1 : Thresholds for constant vertical spacing in the ROAM specification*

Examples of vertical discretisation for various $D_{max}$ are displayed in Figure 3.2.
3.4 Time steps

Time-steps were estimated using CFL condition based on horizontal discretisation, gravity-wave speed (2D time-steps) and internal-wave speed (3D time-steps), assuming constant buoyancy frequency. A safety factor of 0.4 was applied to the CFL condition, creating very conservative estimates of time-steps. The 2D mode time-step is given by:

$$\Delta t_{2D} \leq \frac{0.4}{C_g} \left[ \frac{1}{h_1^2} + \frac{1}{h_2^2} \right]^{-\frac{1}{2}}$$

where:

$$C_g = 2 \sqrt{ghU_{\text{max}}}$$

is the sum of the shallow-water gravity-wave speed plus a nominal maximum expected water velocity ($U_{\text{max}}= 2\text{ms}^{-1}$). Note the factor of 2 is required for the leapfrog time-stepping scheme. The CFL condition for the baroclinic mode is less restrictive, because of the slower speeds of the internal waves, and is given by:

$$\Delta t_{3D} \leq \frac{0.4\beta}{C_i} \left[ \frac{1}{h_1^2} + \frac{1}{h_2^2} \right]^{-\frac{1}{2}}$$

where $C_i = 2\alpha c_i + u_{\text{max}}$ The horizontal phase speed of the fastest internal-wave mode, for the case of continuous stratification with constant buoyancy frequency, $N = (-g\rho_o^{-1}d\rho_o/dz)^{1/2}$, is given by (Gill, 1982, p159):
\[ c_i = \frac{NH}{\pi} = \frac{gH}{\rho_o \pi} \frac{\partial \rho_o}{\partial z} \quad \text{(3.7)} \]

The fastest typical internal wave speeds for the ocean are 2 to 3 m s\(^{-1}\). If the water column is unstratified, \(c_i\) is set to 1 in 3.6. The wave speed derived from Eqn. 3.7 is scaled by a factor of \(\alpha = 2\) before insertion into 3.6, providing a safety net for wave-speed approximation. A further scaling factor of \(\beta = 0.35\) is applied to the baroclinic CFL condition if the maximum depth in the domain is greater than 3000m, and \(\beta = 0.5\) for maximum depths between 1000 and 3000m. Maximum depths less than 1000m use \(\beta = 1\). The time-steps are set to the minimum CFL condition computed at every grid node in the domain.

### 3.5 Horizontal diffusion

The parameterisation for horizontal diffusion could be either explicitly defined, or specified as a function of time and space using the formulation of Smagorinsky (1963). Explicit diffusion was based on the estimation used by Brettschneider (1967) to achieve stability in a model using the Fischer (1959) three time-level scheme;

\[ A_H = 0.01 \frac{h_m^2}{\Delta t_{3D}} \quad \text{(3.8)} \]

where \(h_m\) is the mean grid spacing in either the \(e_1\) or \(e_2\) direction. This diffusion coefficient is limited by the bounds:

\[ 0.05A_{\text{max}} < A_H < A_{\text{max}} \quad \text{(3.9)} \]

where \(A_{\text{max}}\) is the maximum diffusion represented by the CFL condition (Eqn 3.1), and the factor 0.05 was derived on a trial-and-error basis. The horizontal diffusion derived from Eqn 3.8 is scaled to the grid size at every node, that is, for the \(e_1\) direction, \(A_{Hij} = A_H^{*} h_1 / h_m\).

### 3.6 Vertical diffusion

The mixing scheme of Mellor and Yamada (1982) using the triangular analytical formulation for mixing length-scale was used. This scheme appeared more robust in simulations, presumably because it entails no advection of turbulent kinetic energy or dissipation. Background diffusivities and viscosities were set to \(1 \times 10^{-5}\) m\(^2\) s\(^{-1}\). The surface roughness length-scale was set to 1 m. The bottom roughness length was inversely calculated to provide a mean roughness of 0.003, i.e;

\[ z_o = 0.5 \Delta z_{\text{bot}} \left( \exp(k \sqrt{1/0.003}) - 1 \right) \quad \text{(3.10)} \]

where \(k = 0.4\) is Von Karmans constant and \(\Delta z_{\text{bot}}\) is half the bottom layer thickness.
3.7 Advection schemes

The advection scheme of Van Leer (1979) was used for the advection of tracers. This scheme is a higher-order upwind scheme that exhibits positive-definite and monotonic behaviour. Numerical diffusion and dispersion errors produced by the scheme are small. The computational cost of the scheme is fair, resulting in a good all-round advection scheme. A second-order centered scheme was used for momentum advection.

3.8 Initialisation and spin-up

Sea-level, and temperature and salinity distributions were initialised using global model output. Velocity fields started from rest. A ramp of one day was applied to the wind-stress forcing and tidal component of the surface elevation boundary forcing. The low-frequency sea-level component, derived from the global model, did not undergo ramping, to prevent large discontinuities across the open boundaries. The model spun up in a few days. It is expected that the barotropic component would spin up rapidly, but the baroclinic component at depth may take longer. Velocity components were analysed in a profile taken in the Java Trench (Figure 3.3) from a simulation performed in the region south of Java. The maximum depth in this domain was 6717m, one of the deepest regions simulated. Velocity solutions in the $e_1$ and $e_2$ directions at the surface and bottom are displayed in Figures 3.4 and 3.5. Current speeds are displayed in Figure 3.6 and temperature and salinity at the bottom in Figure 3.7. Momentum tendencies for $u_1$ and $u_2$ velocity are displayed in Figures 3.8 and 3.9 respectively. From these figures it can be seen that all variables spin up within 2 days. Furthermore, a simulation was performed for the month of Jan 1999 for the SW Australian region, and a second simulation was re-started from rest at 26 Jan 1999 using the first run for initialisation. Solutions at the end of the month (31st Jan 1999) were compared. Bottom density is displayed in Figure 3.10, from which it is observed that, after 5 days, there is effectively no differences between the simulation started from rest and that spun up for one month.
Figure 3.3. : Bathymetry for the model domain in the Java region. The red dot indicates the site for time series shown in subsequent figures.

Figure 3.4 (a). : Surface velocity component $u_1$ for the first 4 days of the Java Trench model run.

Figure 3.4 (b). : Surface velocity component $u_2$ for the first 4 days of the Java Trench model run.
Figure 3.5 (a). $u_1$ velocity component at 6330m depth

Figure 3.5 (b). $u_2$ velocity component at 6330m depth.

Figure 3.6 (a). 3D current speed at 6330m depth

Figure 3.6 (b). 2D current speed
Figure 3.7 (a). Temperature at 6330m depth

Figure 3.7 (b). Salinity at 6330m depth

Figure 3.8: $u_1$ velocity tendencies (ms$^{-1}$) at 6330m depth

(a) Advection
(b) Horizontal diffusive
(c) Barotropic pressure
(d) Baroclinic pressure
(e) Vertical diffusion
(f) Coriolis
(a) Advevtive
(b) Horizontal diffusive
(c) Barotropic pressure
(d) Baroclinic pressure
(e) Vertical diffusion
(f) Coriolis

Figure 3.9: $u_2$ velocity tendencies (ms$^{-1}$) at 6330m depth

(a) Started 1 Jan 1999
(b) Re-started 26 Jan 1999

Figure 3.10: Comparison of bottom density on the 31 Jan 1999 in SW Australian region for model runs started 26 days apart.
3.9 Relaxation

Temperature and salinity solutions were relaxed to the global T/S using a relaxation timescale of 20 days. This ensured that the density field did not excessively drift from the background state as defined by the global model.
4 Constraining methods

Unconditional stability, requiring that every simulation remain stable, was found to be an unrealistic expectation for a fixed, pre-specified model parameterisation. Unconditional stability could only be achieved through a strategy of combined targeted containment and robust parameterisation. The technique of constraining variables to pre-defined values via nudging or resetting is commonly used in hydrodynamic models, and forms the basis of data assimilation. Targeted filtering of prognostic variables has been used by Klinger et al. (2006) to suppress grid-point noise in velocity fields with the Shapiro filter. Restoring surface temperatures to target observations is a method of approximating surface heat fluxes (Haney, 1971). The practice of relaxing model solutions to climatology to prevent solutions drifting over lengthy simulations is a further example of constraint commonly used (e.g. Hibler and Bryan, 1987). Also, adjustment of 3D velocity is routinely performed to ensure the vertical integral of 3D velocity is identical to the 2D velocity. Therefore, the practice of altering prognostic variables from values provided by the primitive equations is in itself not a new concept. In principle any prognostic variable (e.g. $u_1$, $u_2$, $U_1$, $U_2$, $T$, $S$ or $\eta$), or even tendencies of the momentum balance (advection, barotropic pressure etc) may be constrained. In the present application, it was established that model velocity (2D or 3D) and elevation were required to be selectively constrained when their values exceeded certain predefined thresholds. Additionally, horizontal velocity shear was monitored for reasons discussed in Section 5. The question is: what are the appropriate thresholds on the velocity and elevation, and what action is to be taken if these thresholds are exceeded?

The objective of setting the model thresholds was to find an upper bound of velocity or elevation for ‘realistic’ flows, and also to establish numbers indicative of the model’s approach to instability. Velocities greater than the threshold may indicate a valid transient adjustment process. However a conservative approach was taken in defining thresholds.

An ensemble of model simulations was created by applying the ROAM ocean model to 95 regions around Australia and Indonesia. These regions are illustrated in Figure 4.1. Simulations were generally repeated on each region for January and July 1999, providing a total of 170 different simulations. The simulations were performed using NCEP wind stress rather than output from RAMS. Maximum values of 2D and 3D velocity and elevation were recorded for each simulation, and this dataset formed the basis for establishing thresholds. Of all simulations, 130 (76%) initially produced a stable run. Computed statistics which were used to set the thresholds were confined to the set of stable simulations.

Of the 40 simulations that were not included in the constraint statistics (i.e. considered ‘unstable’) 32 of these (80%, or 19% of total simulations performed) produced a model crash (defined by sea level > 10m). The remaining 8 simulations (20% of ‘unstable’ simulations, or 5% of total simulations) had maximum velocities that were deemed too unrealistic to be included in the threshold statistics. Table 4.1 displays details relevant to these unrealistic solutions.

There were a number of simulations included in the threshold statistics that could be considered ‘dubious’. Although the water speeds would be considered unrealistic for flows in the ocean, they can be considered plausible in the model under conditions of transient adjustment to an impulsive forcing. The cut-off of these ‘dubious’ velocities is somewhat subjective; in this case any velocity with a magnitude greater than 6 ms$^{-1}$ was considered ‘unrealistic’. The simulation statistics based on this ensemble are displayed in Table 4.2. It is acknowledged that the 6 ms$^{-1}$ cut-off may subtly affect the outcome of the threshold statistics (e.g. using the velocity thresholds in Table 4.2 instead of the 6 ms$^{-1}$ cut-off results in new 95-percentile threshold values of $v_{3D} = 3.04$ ms$^{-1}$, $v_{2D} = 2.22$ ms$^{-1}$, $\eta = 0.53$ m).
**Table 4.1 : Details of ‘unrealistic’ simulations classified as unstable**

<table>
<thead>
<tr>
<th>Domain</th>
<th>$v_{3D}$ : Maximum 3D Velocity (ms$^{-1}$)</th>
<th>$v_{2D}$ : Maximum 2D Velocity (ms$^{-1}$)</th>
<th>Maximum $\eta$ difference (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Broome</td>
<td>8.14</td>
<td>2.27</td>
<td>0.61</td>
</tr>
<tr>
<td>Queensland</td>
<td>10.78</td>
<td>4.40</td>
<td>0.39</td>
</tr>
<tr>
<td>Tasman sea</td>
<td>13.31</td>
<td>2.13</td>
<td>0.11</td>
</tr>
<tr>
<td>Kakadu</td>
<td>11.32</td>
<td>13.64</td>
<td>0.69</td>
</tr>
<tr>
<td>EAC</td>
<td>17.0</td>
<td>2.10</td>
<td>0.48</td>
</tr>
<tr>
<td>Kingsound : Jul</td>
<td>30.50</td>
<td>2.99</td>
<td>0.59</td>
</tr>
<tr>
<td>Kingsound : Jan</td>
<td>35.87</td>
<td>3.27</td>
<td>0.65</td>
</tr>
<tr>
<td>Java</td>
<td>58.39</td>
<td>2.73</td>
<td>0.22</td>
</tr>
</tbody>
</table>

* Elevation difference is the difference between the running mean of elevation (i.e. elevation with the tide removed) and the global model elevation.

Since 5% of total simulations were clearly unrealistic, the 95 percentile velocity and elevation values were chosen from the simulation ensemble as the model thresholds. A shear threshold was also established, using the median value because shear was indicative of the presence of $2\Delta x$ noise which had to be suppressed more often (by increasing horizontal diffusion: see Section 5).

The 5% of simulations in Table 4.2 with ‘dubious’ velocities, that is, velocities greater than the threshold, do not yet constitute a large enough sample to allow generalisations on the effect of constraint on an otherwise stable, plausible simulation. The constraint definition will be subject to refinement as further experience is gained in the application of targeted constraint.
<table>
<thead>
<tr>
<th>Variable</th>
<th>Statistic</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Velocity (ms(^{-1}))</td>
<td>Mean</td>
<td>1.67</td>
</tr>
<tr>
<td></td>
<td>Standard deviation</td>
<td>0.87</td>
</tr>
<tr>
<td></td>
<td>Median</td>
<td>1.51</td>
</tr>
<tr>
<td></td>
<td>90(^{th}) percentile</td>
<td>2.67</td>
</tr>
<tr>
<td></td>
<td>95(^{th}) percentile</td>
<td>3.27</td>
</tr>
<tr>
<td></td>
<td>Maximum</td>
<td>5.87</td>
</tr>
<tr>
<td>3D Velocity (ms(^{-1}))</td>
<td>Mean</td>
<td>1.99</td>
</tr>
<tr>
<td></td>
<td>Standard deviation</td>
<td>0.95</td>
</tr>
<tr>
<td></td>
<td>Median</td>
<td>1.83</td>
</tr>
<tr>
<td></td>
<td>90(^{th}) percentile</td>
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</tr>
<tr>
<td></td>
<td>95(^{th}) percentile</td>
<td>3.87</td>
</tr>
<tr>
<td></td>
<td>Maximum</td>
<td>5.87</td>
</tr>
<tr>
<td>2D Velocity (ms(^{-1}))</td>
<td>Mean</td>
<td>1.35</td>
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<tr>
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</tr>
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</tr>
<tr>
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<td>90(^{th}) percentile</td>
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</tr>
<tr>
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<td>2.55</td>
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<tr>
<td></td>
<td>Maximum</td>
<td>4.41</td>
</tr>
<tr>
<td>Elevation difference* (m)</td>
<td>Mean</td>
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</tr>
<tr>
<td></td>
<td>Standard deviation</td>
<td>0.44</td>
</tr>
<tr>
<td></td>
<td>Median</td>
<td>0.28</td>
</tr>
<tr>
<td></td>
<td>90(^{th}) percentile</td>
<td>0.54</td>
</tr>
<tr>
<td></td>
<td>95(^{th}) percentile</td>
<td>0.60</td>
</tr>
<tr>
<td></td>
<td>Maximum</td>
<td>1.73</td>
</tr>
<tr>
<td>Shear** (s(^{-1}))</td>
<td>Mean</td>
<td>9.00x10(^{-4})</td>
</tr>
<tr>
<td></td>
<td>Standard deviation</td>
<td>3.08x10(^{-3})</td>
</tr>
<tr>
<td></td>
<td>Median</td>
<td>2.03x10(^{-4})</td>
</tr>
<tr>
<td></td>
<td>90(^{th}) percentile</td>
<td>9.31x10(^{-4})</td>
</tr>
<tr>
<td></td>
<td>95(^{th}) percentile</td>
<td>1.47x10(^{-3})</td>
</tr>
<tr>
<td></td>
<td>Maximum</td>
<td>1.64x10(^{-2})</td>
</tr>
</tbody>
</table>

* Elevation difference is the difference between the running mean of elevation (i.e. elevation with the tide removed) and the global model elevation.

** Based on 57 simulations.

* Table 4.2 : Statistics for stable simulations. Text in orange are the values used as ROAM thresholds.
The methods of constraint were as follows:

**Velocity**

A nine point (3x3) median filter was used to establish a new velocity for those cells where velocity exceeded the threshold. The median filter is well known for its use in the suppression of noise, and is considered better than a mean filter, since any single unrepresentative value in the nine-point stencil will not affect the median value significantly. In the case where the median also exceeded the threshold, velocity was constrained to the threshold value. Resetting velocity will not impact on the conservation characteristics of the model if it is done at the correct point in the sequence of computations. For the 2D mode, the velocity at the current time-step is constrained directly after the Asselin (1972) time-filtering (for leapfrog time-stepping), and before the elevation is updated and velocities are stepped forward in time. Any constrained velocities are then guaranteed to be used in the elevation calculation. For the 3D mode, updated velocities are constrained after the time-filtering and stepping forward in time, but before the vertical integral of 3D velocity is adjusted to the 2D velocity. This ensures that fluxes for tracer computation are using constrained, adjusted 3D velocities. The procedure of resetting velocity may be considered as imposing an additional source or sink in the momentum balance. Alternatively, drag may be altered to achieve the desired result, or horizontal viscosity may be inversely computed to achieve the reset velocity.

**Elevation**

The practical use of the ensemble elevation maximum is problematic, since any threshold must encompass the natural (large) tidal variability, yet discriminate against elevation increases due to instability. Hence the tidal signal is removed to establish whether the background elevation is unduly increasing. The tide is removed from the sea-level solutions by computing a running average of sea-level. As an alternative, the global tide model could be used to estimate and remove the tide at every grid node. However, this method proved generally less reliable than using a running mean, especially in shallow coastal areas where the quality of the global tide model prediction deteriorates.

The tidal removal also allows direct comparison to the global model output. The model mean elevation, or low-frequency sea-level component, was relaxed to the global model elevation when the difference between the mean elevation from the simulation and the global model simulation exceeded the threshold. Relaxation to the global model elevation was performed at a timescale proportional to the difference in low frequency elevations when thresholds are exceeded, e.g.

\[
rate = 20\Delta t_{2D} \quad \text{if} \quad \Delta \eta = \eta_T
\]

\[
rate = 2\Delta t_{2D} \quad \text{if} \quad \Delta \eta \geq 2\eta_T
\]

4.1

where \(\Delta \eta\) is the difference in low frequency elevations and \(\eta_T = 0.6\) m is the elevation threshold.

Relaxing elevation will render the model un-conservative unless a flux of tracer is added to compensate for the changed volume of surface cell in the tracer equations. Unfortunately there is no analogous relaxation procedure available for velocity due to the inclusion of the tidal component.
5 Discussion

Of all simulations performed, 76% executed cleanly with no requirement for constraint, indicating that the ROAM parameterisation is relatively robust. The remaining simulations were required to be constrained although, for the majority of these, constraint was only necessary for less than 1% of the simulation, and only at a few cells. Most simulations requiring containment were in regions subject to complex large-scale ocean dynamics (e.g. western boundary currents, or the Indonesian throughflow). Although all constrained simulations were stable, some solutions exhibited inaccuracies, particularly in velocity solutions adjacent to open boundaries near coastal margins. Occasionally a checkerboard pattern due to nonlinear instability ($2\Delta x$ noise) was evident in the sea level ($\eta$) solution. This is attributed to insufficient horizontal diffusion generated from the Smagorinsky scheme (e.g. Haidvogel and Beckmann, 1999, with more references in Griffies et al., 2000, p136). When constant explicit diffusion was used in the model these inaccuracies in $\eta$ were removed, at the expense, however, of detailed structure in the temperature and salinity solutions. This is demonstrated in Figure 5.1 and 5.2. Figure 5.1 (a) shows the $\eta$ solution for a region surrounding the south of the island of Celebes using the standard ROAM parametrisation. Figure 5.2 (a) uses explicit horizontal diffusion of $7.4 \times 10^3$ m$^2$s$^{-1}$. This second formulation removes the checkerboard in $\eta$, but structure in the temperature solution is considerably reduced (Figures 5.1 b and 5.2 b). Mellor and Blumberg (1985) have also noted that small diffusion leads to noisy solutions, while large values remove resolvable flow structures. Increasing the empirical coefficient in the Smagorinsky scheme to 0.5 (default is 0.1) and smoothing the resultant viscosity using a 9 point convolution filter also stabilizes the $\eta$ solution (Figure 5.3 a) whilst preserving structure in the temperature solution (Figure 5.3 b). Hall and Davies (2005) noted that increasing the empirical Smagorinsky coefficient resulted in excessive smoothing of internal wave solutions on an irregular grid. However, these authors found an optimum value of the constant that resulted in a solution without excessive smoothing whilst maintaining stability. Experience is still required in this area to determine the optimum constant in the Smagorinsky scheme.

![Surface elevation](image1.png) ![Surface temperature](image2.png)

(a) : Surface elevation  (b) : Surface temperature

Figure 5.1 : Southern Celebes simulation using the standard ROAM Smagorinsky diffusion.
(a) Surface elevation  
(b) Surface temperature  

Figure 5.2: Southern Celebes simulation with a constant diffusion of 7.4x10^3 m^2 s^-1.

(a) Surface elevation  
(b) Surface temperature  

Figure 5.3: Southern Celebes simulation using a Smagorinsky diffusion coefficient of 0.5 (compared with 0.1 in Fig. 5.1).
The relatively small viscosities generated by the Smagorinsky scheme can also result in $2\Delta x$ noise in the velocity solution (Figure 5.4 a). This effect can again be removed using larger constant horizontal viscosities (Figure 5.4 b). As mentioned in section 4, velocity shear (which is indicative of $2\Delta x$ noise) was monitored and, when shear exceeded the prescribed threshold, horizontal viscosity was increased locally. This strategy performed well in eliminating $2\Delta x$ velocity noise; Figure 5.5 (a)-(d) shows the $u_2$ velocity component, horizontal viscosity and T/S solutions for a region surrounding Brisbane. The figure clearly shows the error in the velocity field, but error is not apparent in the T/S solutions. Increasing viscosity locally (Figure 5.6 b) when shear increases above the threshold of $2.03 \times 10^{-4} \text{ s}^{-1}$ removes this velocity error (Figure 5.6 a). T/S solutions remain largely unaffected by the increased viscosity (Figures 5.6 c & d).

![Figures 5.4 and 5.6](image_url)

**Figure 5.4**: Comparison of $u_1$ velocity component around New Britain for simulations with different horizontal diffusion.

(a) Smagorinsky diffusion, coefficient = 0.1

(b) Constant diffusion of $7.4 \times 10^3 \text{ m}^2 \text{s}^{-1}$
Figure 5.5: Modelled surface solutions in the Brisbane region showing instability in the $u_2$ signal.
(a) $u_2$ velocity component (ms$^{-1}$) 

(b) Viscosity in the $e_2$ direction (m$^2$s$^{-1}$)

Figure 5.6: Model solutions in the Brisbane region, as in Fig. 5.4, but with enhanced viscosity
In certain instances, unrealistically large velocities appeared adjacent to the coast on cross-shelf boundaries. These typically occurred at the upstream boundary (i.e. the boundary experiencing inflow) and are attributed to boundary over-specification issues related to the deteriorating accuracy of the global tide model in shallow water. This phenomenon is illustrated in Figure 5.7, which shows large surface velocities adjacent to the coast on the northern boundary in a simulation of the East Australian Current region. These large velocities had minimal impact on temperature and salinity solutions.

![Figure 5.7: Surface velocity and temperature in the East Australian Current region.](image)

T/S solutions generally exhibited no obvious source of error. However, recirculations at the open boundary were occasionally evident. Figure 5.8 shows example recirculations in the velocity field in the Tasman Sea. Such features were typically observed in mid-ocean simulations containing four open boundaries. Occasionally the recirculations had a tendency to propagate in an anti-clockwise direction around the open boundary (i.e. with the open boundary on the right relative to the direction of propagation). There were also instances where the recirculations appeared on cross-shelf open boundaries adjacent to land. When these recirculations were large, the T/S solutions at the open boundaries exhibited an excessive influx of heat or salt. Figure 5.9 (a) shows a velocity and temperature solution in the Brisbane region when a recirculation is present on the southern open boundary. The temperature solution shows a plume of lower temperature water propagating into the domain adjacent to the coast as a result of the spurious velocities. Figure 5.9 (b) shows the corresponding global model temperature solution for reference. Generally it is unclear in the model solutions if an inflow, or recirculation, at the open boundary is a result of the propagation of a legitimate feature into the domain (e.g. an eddy) or a boundary artifact. For example, Figure 5.10 (a) shows surface currents and temperature for the west coast of Australia, where there is an apparent inflow of warm water on the northern boundary due to a boundary recirculation. From the global model temperature solution (Figure 5.10 b), this feature is revealed as the propagation into the domain of a warm instability associated with the weak summer Leeuwin Current. Cool features on the boundary in the south west of the domain also are evident in the global model solution.
Figure 5.8: ROAM velocity and temperature in the Tasman Sea showing boundary recirculation features.

Figure 5.9: Solutions in the Brisbane region, with ROAM showing boundary recirculations in the velocity and temperature field.

(a) ROAM surface velocity and temperature

(b) OFAM surface temperature (°C)

Figure 5.9: Solutions in the Brisbane region, with ROAM showing boundary recirculations in the velocity and temperature field.
The sponge scheme applied at the open boundaries was on occasion responsible for a smearing of the T/S solutions. This is demonstrated in Figure 5.11 in a simulation of the region around Ceduna, SA, where the temperature solution exhibits a ‘stripe’ devoid of structure adjacent to the open boundary. Note that sponges are only applied to horizontal viscosity at the boundary; horizontal diffusion for tracers is not subjected to increased diffusivity. These ‘stripes’ are therefore a result of the large viscosity homogenising the velocity field near the boundaries, and smearing eddy structure, resulting in relatively uniform influx of heat and salt across the boundary. Error in the temperature and salinity solutions as a result of boundary smearing or re-circulations was evident in only a small percentage of the total simulations.
Temperature and salinity fields from ROAM differed from the global model solution in ways that depended on the application. Often the global model and ROAM exhibited similar solutions (e.g. Figure 5.12 displaying the temperature solution for Torres Strait). In certain regions, the ROAM T/S solutions showed more realistic structure than the global model (e.g. Figure 5.13 for Port Phillip Bay, resolution = 0.01°; Figure 5.14 for the Christchurch region, south NZ, resolution = 0.02°). These instances were typically associated with higher resolution domains. Various domains showed T/S solutions markedly different from the global model. These were typically in regions of complex flow, especially Indonesia, where the pressure-forced open boundary communicated poorly the general circulation into the regional domain (e.g. Figure 5.15, East Timor region, resolution = 0.06°). Occasionally the global model exhibited more structure (Figure 5.16, Vanuatu region, resolution = 0.08°). The T/S solutions varied in regions where local atmospheric heat or freshwater exchanges were important (Figure 5.17, SA Gulfs, resolution = 0.04°). (ROAM was executed without heat or salt fluxes due to a lack of consistently reliable data).

Figure 5.12: Surface temperature (°C) in the Torres Strait region from both ROAM and OFAM.
Figure 5.13: Surface temperature (°C) in the Port Phillip Bay region, showing the effect of higher resolution in ROAM compared with OFAM.

Figure 5.14: Surface temperature (°C) in the Christchurch region, showing the effect of higher resolution in ROAM compared with OFAM.
Figure 5.15: Surface temperature (°C) in the East Timor region, where open-boundary conditions appear to have compromised the ROAM solution.

Figure 5.16: Surface temperature (°C) in the Vanuatu region, where open-boundary conditions appear to have compromised the ROAM solution.
ROAM and the global model are forced differently. ROAM includes tides whereas the global model did not. This creates the potential for altered current regimes through tidally rectified mean flows and increased bottom friction. This difference is especially relevant where tides are large, as in northern Australia. Further, the level-2.0 turbulence-closure scheme used in ROAM will create vertical mixing regimes differing from those in the global model. As in Figure 5.17, there are instances when surface heat and freshwater fluxes may cause ROAM and global T/S solutions to diverge. Finally, the boundary forcing used in ROAM consisted only of sea-level and T/S profiles (i.e. pressure). In regions of non-linear behaviour (boundary currents, Indonesian Throughflow), such forcing may be insufficient to communicate basin-scale phenomena accurately into the ROAM domain. In these cases, the boundary forcing in ROAM is under-specified and the solutions may diverge from OFAM. It would be preferable to force open boundaries with velocity solutions from the global model but this is not possible, because the global model does not include tides.

The observed differences in global model and ROAM solutions are attributed to a combination of the above factors. The ROAM and OFAM simulation comparisons presented here were from simulations of one month duration. Differences in T/S solutions after a one-week simulation were small.

The resolution of the global model is 0.1° (~11 km) in the Australasian region. Idealised experiments using various nesting ratios (Spall and Holland, 1991) revealed that ratios of coarse to fine grid resolutions of 3:1 and 5:1 produce acceptable results, whereas ratios of 7:1 began to introduce error. The ratio of 5:1 (i.e. down to 0.02°) is therefore considered a practical limit for nesting ratios. A histogram detailing the range of resolutions represented by the simulation ensemble is presented in Figure 5.18, which indicates that a number of
simulations ran successfully at ratios higher than 5:1 (the highest-resolution simulation was 0.005°). However, there is no measure of the accuracy of these simulations. The higher-resolution simulations typically exhibited enhanced structure in T/S solutions compared to the global model. Further investigation is required to assess the deterioration of model results at nesting ratios beyond 5:1. The largest number of simulations was performed with resolution of 0.04°, corresponding to a nesting ratio of approximately 2.5:1.

![Histogram of simulation resolutions](image-url)

*Figure 5.18: Histogram of simulation resolutions.*
6 Conclusions

Unconditional model stability has been achieved in ROAM with a fixed parameterisation (based on the user-specified geographic bounds and resolution of a model domain, and a start / stop time), in conjunction with a strategy of constraint. Stable simulations were produced for 130 runs encompassing a range of resolutions, geographies, forcing regimes and dominant physics. In the majority of simulations, the model solutions were free from obvious numerical error, and temperature and salinity solutions exhibited general features similar to the global model. ROAM T/S solutions generally displayed more structure than OFAM T/S, especially as resolution increased.

Constraint of velocity and elevation may undesirably perturb the model solutions. However the consequence of instability is total failure of the model simulation. Although the ROAM methodology has achieved unconditional stability, the solutions cannot be considered unconditionally accurate. Occasionally, the model must adaptively shift to the robust end of the robustness - accuracy spectrum, to preserve stability, with associated consequences for accuracy. Numerical artifacts were occasionally obvious in model solutions, although these are more often associated with the velocity field than the temperature or salinity.

The accuracy and consistency of ROAM are assessed by comparison with data and optimised models in the accompanying report (Herzfeld and Cahill, 2006).
7 References


