Modeling studies of barotropic and baroclinic dynamics in the Malacca Strait

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Abstract. The circulation in the Malacca Strait is simulated using a three-dimensional barotropic numerical model. The circulation model is derived from the combined effects of tides, wind, meteorological forcings, temperature and salinity. The computational results produced pattern of general circulation and also sea surface temperature and salinity. In the present study, the pattern of current circulation in the Malacca Strait coincided with the...
2. HAMSOM Model

The continuous partial differential equations of the model are as follows:

\[
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - fv = -\frac{1}{\rho} \frac{\partial p}{\partial x} + \frac{\partial}{\partial y} \left( A_H \frac{\partial u}{\partial y} \right) + \frac{\partial}{\partial z} \left( A_V \frac{\partial u}{\partial z} \right) \tag{2}
\]

Y-component momentum equation:

\[
\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + fu = -\frac{1}{\rho} \frac{\partial p}{\partial y} + \frac{\partial}{\partial x} \left( A_H \frac{\partial v}{\partial x} \right) + \frac{\partial}{\partial z} \left( A_V \frac{\partial v}{\partial z} \right) \tag{3}
\]

The variables are the three components of the velocity \( u, v, \) and \( w \), the pressure \( p \), the density \( \rho \), the three space variables, i.e. \( x \) (positive in the east direction), \( y \) (positive in the north direction), \( z \) (positive upwards) and of the time \( t \), the Coriolis acceleration \( f \). The variables \( A_H \) and \( A_V \) are the horizontal and vertical coefficients of turbulent viscosity, \( F_x \) and \( F_y \) are the components of the horizontal exterior forces.

Hydrostatic equation:

\[
\frac{\partial \rho}{\partial z} = -\rho g \tag{4}
\]

Where \( g \) is the acceleration due to gravity.

The continuity equation (1) is integrated over an arbitrary vertical model layer \( h \);
\[ \Delta \omega = \int_{z}^{z+h} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) dz \]

\[ = -\frac{\partial}{\partial x} \left( \int_{z}^{z+h} u dz \right) - \frac{\partial}{\partial y} \left( \int_{z}^{z+h} v dz \right) + u_{z_0} \frac{\partial}{\partial x} (z_0 + h) - v_{z_0} \frac{\partial}{\partial y} (z_0 + h) + w_{z_0} \frac{\partial}{\partial z} \]

\[ = -u \frac{\partial h}{\partial x} - v \frac{\partial h}{\partial y} + u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y} \]

The layer thickness \( h \) is constant except the top and bottom layers, so for the layers within the water column (5) is reduced to:

\[ \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} + \Delta \omega = 0 \] (6)

For the top layer:

\[ \frac{\partial h}{\partial x} = \frac{\partial \zeta}{\partial x} \quad \text{and} \quad \frac{\partial h}{\partial y} = \frac{\partial \zeta}{\partial y} \]

Hence \( \Delta \omega = -\frac{\partial U}{\partial x} - \frac{\partial V}{\partial y} + u \frac{\partial \zeta}{\partial x} + v \frac{\partial \zeta}{\partial y} \)

Neglecting the nonlinear terms, we get:

\[ \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} + \Delta \omega = 0 \] (7)

For the bottom layer:

\[ \frac{\partial h}{\partial x} = \frac{\partial H}{\partial x} \quad \text{and} \quad \frac{\partial h}{\partial y} = \frac{\partial H}{\partial y} \]

hence \( \Delta \omega = -\frac{\partial U}{\partial x} - \frac{\partial V}{\partial y} + u \frac{\partial H}{\partial x} + v \frac{\partial H}{\partial y} \) (8)

By neglecting the nonlinear terms in the surface layer, and taking the zero value for \( \omega \) at the seabed, the general form of the layer-averaged continuity equation is:

\[ \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} + \Delta \omega = 0 \] (9)

The overall continuity equation is obtained by integrating over the entire water column:

\[ \frac{\partial \zeta}{\partial t} + \frac{\partial \bar{U}}{\partial \bar{x}} + \frac{\partial \bar{V}}{\partial \bar{y}} = 0 \] (10)
2.1 Momentum equations
The detailed deduction of the layer-averaged equations of motion is omitted. Interested readers are referred to [6]. The layer-averaged equations of motion are as follows:

X-component equation:
\[
\frac{\partial U}{\partial t} + \frac{\partial (UU / h)}{\partial x} + \frac{\partial (VU / h)}{\partial y} + \Delta (u\omega) - fV = \frac{h}{\rho} \frac{\partial p}{\partial x} + A_h \nabla^2 U + \Delta \tau^\ast + hF_x
\] (11)

Y-component equation:
\[
\frac{\partial V}{\partial t} + \frac{\partial (UV / h)}{\partial x} + \frac{\partial (VV / h)}{\partial y} + \Delta (v\omega) + fU = -\frac{h}{\rho} \frac{\partial p}{\partial y} + A_h \nabla^2 V + \Delta \tau^\ast + hF_y
\] (12)

The shear stress term \( \tau \) are specified at the upper and lower interfaces of the model layer \( h \).

\[
\begin{bmatrix}
\tau^x \\
\tau^y
\end{bmatrix} = A_v \begin{bmatrix}
\frac{\partial U}{\partial z} \\
\frac{\partial V}{\partial z}
\end{bmatrix} = A_v \begin{bmatrix}
\frac{U}{h} \\
\frac{V}{h}
\end{bmatrix}
\] (13)

Where the vertical eddy viscosity \( A_v \) is a function of time and space, depending on both the current shear and the stability of the fluid.

2.2 Equations for temperature and salinity
The layer-averaged equations for temperature and salinity are obtained by taking the structure of temperature and salinity to be a stepwise function, i.e. within the vertical model layer the temperature and salinity are assumed to be uniform.

Temperature conservation equation:
\[
\frac{\partial T}{\partial t} + U \frac{\partial T}{\partial x} + V \frac{\partial T}{\partial y} + \omega \frac{\partial T}{\partial z} = K_T \nabla^2 T + \frac{\partial}{\partial z} \left( K_T \frac{\partial T}{\partial z} \right) + S_T
\] (14)

Salinity conservation equation:
\[
\frac{\partial S}{\partial t} + U \frac{\partial S}{\partial x} + V \frac{\partial S}{\partial y} + \omega \frac{\partial S}{\partial z} = K_S \nabla^2 S + \frac{\partial}{\partial z} \left( K_S \frac{\partial S}{\partial z} \right) + S_S
\] (15)

From the above deduction, the general forms of the governing equations, which are similar to the equations used by [1] are as follows:

The equations of motion are:
\[
\frac{\partial (U / V)}{\partial t} + (U \frac{\partial U}{\partial x} + V \frac{\partial U}{\partial x}) = -\frac{h}{\rho} \left( \frac{\partial p}{\partial x} \right) + X
\] (16)

Where \( X \) are the further optional terms (isolated since they are treated with explicit scheme):
\[
X = -u \frac{\partial U}{\partial y} + v \frac{\partial W}{\partial z} + A \nabla^2 U + hF - \Delta [uw]
\]
\[
Y = -u \frac{\partial V}{\partial y} + v \frac{\partial W}{\partial z} + A \nabla^2 V + hF - \Delta [uv]
\] (17)

The hydrostatic pressure \( p \) (excluding the \( g \rho \zeta \) and \( p \) terms, because they have no contribution to the horizontal pressure gradient) at a depth \( z \) is given by:
\[
p = g \rho \zeta + p^0(\zeta) + \int_0^z p^0 dz = g \rho \zeta + I
\] (18)
Due to the different stability limitations of the time step to the numerical simulations, the pressure in the above equation is separated into two parts, i.e. its barotropic component \((g\rho_1 \zeta)\) and its baroclinic component \(I\).

The equation of continuity is given by (9). The prognostic equation for the sea surface elevation is given by (10).

The equation of temperature is:

\[
\frac{\partial T}{\partial t} + \omega \frac{\partial T}{\partial \zeta} = \frac{\partial}{\partial \zeta} \left( K_V \frac{\partial T}{\partial \zeta} \right) + O^T
\]

(19)

The equation of salinity is:

\[
\frac{\partial S}{\partial t} + \omega \frac{\partial S}{\partial \zeta} = \frac{\partial}{\partial \zeta} \left( K_V \frac{\partial S}{\partial \zeta} \right) + O^S
\]

(20)

Where \(O^T\) and \(O^S\) are the further optional terms of temperature and salinity.

2.3 Numerical scheme

For the approximation of the time domain a two time-level scheme is introduced, denoted by the time-level indices \(n\) and \(n+1\), where the latter index represent the future time-level. In this scheme the prognostic variables \((\zeta, U, V, T, S)\) which enter the semi-implicit algorithms, are defined at staggered time-levels. Pressure is separated into its barotropic component \(g\rho_1 \zeta\), and its baroclinic component \(I(18)\). A semi-implicit scheme is introduced to the barotropic pressure gradient terms in the equations of motion (16), to the horizontal divergence term in the equation of continuity (10), to the vertical shear stress terms in the equations of motion(3.32), and also to the vertical diffusion terms in the equations of temperature and salinity (19, 20).

In the equations of motion (16) a stable second order approximation for the Coriolis terms is introduced, in order to avoid the linear numerical instability arising from a forward-in-time approximation of the Coriolis acceleration.

2.3.1 Model arrangement for the Malacca Strait

The model region covers 95.5 E to 103.5E and 1.5 N to 5.5 N (Fig. 1). In this study, the Malacca Strait is discretized with a horizontal mesh size of \(\Delta x = \Delta y = 5\) angular minutes. In vertical direction, the model has 17 layers.

The time-step is \(\Delta t = 300\) s. At the open boundaries, amplitudes and phases of the five major tidal constituents (M\(_2\); S\(_2\); N\(_2\); K\(_1\); O\(_1\)) are prescribed from a global tidal model (see. [9]) and T and S from climatological data of [5]. The atmospheric forcing, i.e. winds and surface heat fluxes are derived from the NCEP/NCAR reanalysis data [see. 4].

3. Results and Discussions

The surface current circulation is presented in Figs. 2 and 3. The model is derived from the combine effect of tides, wind, heat flux which obtained from NCEP/NCAR reanalysis data.
Figure 2. Surface current in northeast monsoon

Figure 3. Surface current in southwest monsoon
Figure 4. Surface current in southwest monsoon based on [9]

Figure 5. Surface current in northeast monsoon based on [9]
The circulation in the Malacca Strait has an impact to the water mass exchange between South China Sea and Indian Ocean. Generally, the water movements are in general directed towards the Indian Ocean and are strongly related to the surface gradient of the sea level through this strait. In the Malacca Strait, the period of strongest flow is from January to April, during northeast monsoon, however it is chiefly caused by the low sea level in the Andaman Sea in this season.

In northeast monsoon, water mass with high salinity flow from the South China Sea through the Malacca Strait. While in southwest monsoon, the low salinity in the south-western part of the Malacca Strait is caused by the water mass from the Java Sea.

Sea surface salinity is presented in Fig. 8. During southwest monsoon, surface salinity is lower in the northwest of the Malacca Strait. As has been stated above, strong wind from southwest monsoon leads to maximum rainfall over most parts of the Asian subcontinent from June to September. Freshwater from the rivers is largely responsible for lowering the salinity.

Figure 6. Surface current in northeast monsoon based on [3]

Figure 7. Surface current in southwest monsoon based on [3]
Figure 8. Sea surface salinity (January-December), based on HAMSOM Model. Sea surface temperature is presented in Fig. 9. Surface water temperature is warm, ranging from 28°C-30°C and 26°C-28°C during southwest and northeast monsoon, respectively.
Figure 9: Sea surface temperature (January-December), based on HAM-SOM Model.

4. Conclusions
The computational results produced pattern of general circulation and also sea surface temperature and salinity. In the present study, the pattern of current circulation in the Malacca Strait coincided with the work
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