
The Numerical models and the forcing parameters

2.1 Introduction

The motions in the oceans are turbulent and span a wide spectrum of temporal and spatial scales. Oceanic circulation is a function of the internal density structure of the water masses, radiative fluxes at the sea surface and the forcing imposed at the sea surface by the overlying atmosphere, mainly the wind stress and buoyancy fluxes. The surface forcing contains a variety of temporal scales ranging from hourly variations to decades and beyond and spatial scales ranging from kilometers to thousands of kilometers. The oceans respond in a very complicated manner to the atmospheric forcing and we do not have any information on the nature of oceanic response to the wide spectrum of surface forcing mainly because the observations covering the entire spectrum do not exist. Ocean models are the tools to understand the response of the oceans to the wide spectrum of atmospheric variability, to simulate circulation on different time and space scales and to predict the oceanic variability on shorter time scales of few days. Comprehensive ocean models in combination with the sparse insitu and relatively abundant remotely sensed data, provide the best means for studying and monitoring the oceans . The overall structure of the oceans and the broad features of circulation have been obtained by insitu observations; however, observational studies cannot give quantitative information on the nature of circulation and the variability of circulation under different atmospheric conditions.

Different kinds of models are being used for the study of the oceans depending on the intended application and on the computational and the pre- and post-processing capabilities available. Although, all numerical models are based on more or less the same set of governing equations, the performance of each model differs widely. This has to do with the fact

that when the same continuum equations are discretized for solution on a digital computer, the behaviour of the resulting discretised forms of governing equations can differ widely depending on the horizontal and vertical discretization schemes used, time stepping methods and the manner in which certain important terms such as advection and diffusion are handled. The choice of vertical coordinate is an important parameter in ocean modeling because the bottom topography and its gradients are of overwhelming importance to ocean circulation and to the horizontal and vertical transport of properties such as temperature and dissolved substances. At present, no ocean model can cope with the large topographic gradients formed along the continental slope and at many places in the ocean.

Three different types of vertical co-ordinates are currently being used in ocean models; the Z-coordinate model developed at the Geophysical Fluid Dynamics Laboratory (Bryan, 1969; Cox, 1984). The Miami Isopycnal Coordinate Ocean Model (MICOM) (Bleck and Smith, 1990) and the topography following sigma coordinate model developed at Princeton (Princeton Ocean Model, POM) (Blumberg and Mellor, 1987) are the most prominent ocean models in different vertical co-ordinates. The simplest choice of vertical coordinate is Z, which represents the vertical distance from the free ocean surface at $Z=0$, with Z positive upwards, to $Z=-H(x,y)$, the bottom topography. The Z coordinate models have been used for many decades and are still being used (Bryan, 1969; Semtner 1974; Cox, 1984). The principal advantage of the Z coordinate models is that the horizontal pressure gradient terms which are important to the momentum balance can be computed accurately. The equation of state of ocean waters, which is highly non-linear and important for determining water mass properties, can also be represented accurately. In the Z coordinate models, the number of levels varies from grid point to grid point and it depends on the local depth. Unless high vertical resolution is provided in the upper layers, shallow shelf regions cannot be modeled properly.

Hence, Z-coordinate models have been used primarily for basin scale and global simulations.

In the sigma coordinate model, the governing continuity, momentum and scalar conservation equations are cast in a bottom topography following coordinate system by defining a new variable $\sigma = Z/H$. So that $\sigma = -1$ defines the bottom. In this coordinate system, the number of levels are the same everywhere in the ocean irrespective of the depth of the water column. Hence, both the shallow shelf and deep ocean regions can be resolved with out excessive number of vertical levels and the kinematic boundary conditions at the bottom can be satisfied correctly. But the σ -models have difficulties in accurately representing the horizontal pressure gradient because the surfaces of constant sigma are not generally horizontal. The horizontal pressure gradient which is perpendicular to the local vertical direction as defined by gravity will have a projection along and across the sigma surfaces. This results in spurious pressure forces that drive non-trivial unphysical currents. Hence, regions of sharply changing topography such as continental slope are poorly represented in these models.

There have been many studies on the modeling of circulation of the Indian Ocean. A brief description of various modeling studies on the Indian Ocean is given in Chapter 1. But all these modeling studies were concentrated on the application of reduced gravity models or prognostic models based on idealized forcing functions. These existing models mainly deal with upper layer circulation and there is hardly any information on the circulation in the intermediate, deep and bottom layers. The existing numerical models of Indian Ocean also do not incorporate all the site specific forcing functions such as fresh water fluxes and river discharge adjoining the Indian coast line. The main objective of this chapter is to study the 3-dimensional circulation using a 33 level state of art sigma coordinate ocean model that can be run on diagnostic, semi-

diagnostic and prognostic models for the Indian Ocean area to study the dynamics of large scale circulation in the tropical Indian Ocean. In this chapter, the sigma coordinate model equations, boundary and initial conditions and the forcing parameters of the model are presented.

2.2 The numerical models

2.2.1 Model equations, boundary and initial conditions in Z-coordinate

The numerical model used in the present study has been originally developed by Alekseev and Zalesny (1993) at the Institute of Numerical Mathematics, Russian Academy of Sciences, Moscow and the same has been subsequently modified by Zalesny and Galkin (1996) and Bangno et al. (1996a, 1996b). The governing equations of the Ocean General Circulation Model are the primitive equations of motion in spherical coordinates, with hydrostatic, Boussinesq and rigid lid approximations. We first give below the governing equations in Z- coordinate and then the transformed equations in σ -coordinate system are presented, which can be run on diagnostic, semi-diagnostic and prognostic modes. The peculiarity of the σ - coordinate models in diagnostic, semi-diagnostic and prognostic models will be discussed in chapter III and IV where detailed analysis of the results are also presented.

The basic hydrodynamic equations that govern the large scale 3-D flows in Z-coordinate are written as follows (Zalesny et al., 2000);

$$\frac{du}{dt} - (l + m \sin \theta) v = -\frac{m}{\rho_0} \frac{\partial P}{\partial \lambda} + D''u + \frac{\partial}{\partial z} \left(v \frac{\partial u}{\partial z} \right) \quad \text{————— (2.1)}$$

$$\frac{dv}{dt} + (l + m \sin \theta)u = -\frac{n}{\rho_0} \frac{\partial P}{\partial \theta} + D''v + \frac{\partial}{\partial z} \left(v \frac{\partial v}{\partial z} \right) \quad \text{_____ (2.2)}$$

$$\frac{\partial P}{\partial Z} = \rho g \quad \text{_____ (2.3)}$$

$$m \left(\frac{\partial u}{\partial \lambda} + n \frac{\partial}{\partial \theta} \left(\frac{v}{m} \right) \right) + \frac{\partial w}{\partial z} = 0 \quad \text{_____ 2.4)}$$

$$\frac{dT}{dt} = \tilde{D}^T T + \frac{\partial}{\partial z} \left(v_r \frac{\partial T}{\partial z} \right) \quad \text{_____ (2.5)}$$

$$\frac{dS}{dt} = \tilde{D}^S S + \frac{\partial}{\partial z} \left(v_s \frac{\partial S}{\partial z} \right) \quad \text{_____ (2.6)}$$

$$\rho = f(T, S, P) - \rho_0 \quad \text{_____ (2.7)}$$

In the above equations (2.5), the parameter T represents the potential temperature and not insitu temperature. In equation (2.7), the effect of compressibility is taken into consideration (Bryden et al, 1999). Equations (2.1)-(2.7) are solved subject to the following boundary and initial conditions.

At the surface, a rigid lid boundary condition for vertical velocity was employed. Specification of rigid lid boundary condition at the surface will eliminate short period surface gravity waves from the solution and it will enable us to use larger time steps of integration. However, large scale waves such as Rossby and Kelvin waves are retained in the solution. The wind stress at the surface is expressed in terms of the gradient of velocity in the vertical multiplied by the vertical turbulent mixing coefficient and

mean density of sea water. Thus at the surface, ($z = 0$), the following boundary conditions are specified.

$$W_{z=0} = 0 ; \quad v \frac{\partial u}{\partial z} = -\frac{\tau^\lambda}{\rho_0} ; \quad v \frac{\partial v}{\partial z} = -\frac{\tau^\theta}{\rho_0} \quad \text{_____ (2.8)}$$

The boundary conditions for temperature and salinity are ;

$$\beta_T v_T \frac{\partial T}{\partial z} + \alpha_T T = Q_T ; \quad \beta_S v_S \frac{\partial S}{\partial z} + \alpha_S S = Q_S \quad \text{_____ (2.9)}$$

Depending on the values of the coefficients $\alpha_T, \beta_T, \alpha_S$ and β_S , the above mentioned boundary conditions will be transformed into either Dirichlet boundary condition or Neuman boundary condition.

At the bottom ($Z = H(\lambda, \theta)$), the following boundary conditions were specified.

$$w = mu \frac{\partial H}{\partial \lambda} + nv \frac{\partial H}{\partial \theta} ; \quad \frac{\partial u}{\partial z} = \frac{\tau_b^\lambda}{\rho_0} ; \quad \frac{\partial v}{\partial z} = \frac{\tau_b^\theta}{\rho_0} \quad \text{_____ (2.10)}$$

For temperature and salinity, the following boundary conditions were specified.

$$\frac{\partial T}{\partial N_T} = 0 ; \quad \frac{\partial S}{\partial N_S} = 0 \quad \text{_____ (2.11)}$$

Along the rigid lateral boundary of the model the normal flux boundary condition of zero velocity is prescribed.

$$u = v = 0$$

For temperatures and salinity, the following condition was applied.

$$\frac{\partial T}{\partial N_T} = 0 \quad ; \quad \frac{\partial S}{\partial N_S} = 0 \quad \text{_____} (2.12)$$

where N is the normal to the Lateral boundary. The initial conditions are

$$u = u^0; v = v^0; T = T^0; S = S^0 \quad \text{_____} (2.13)$$

In the above mentioned equations (2.1) - (2.13), meanings of the symbols are as follows :

$$\frac{d}{dt} = \frac{\partial}{\partial t} + mu \frac{\partial}{\partial \lambda} + nv \frac{\partial}{\partial \theta} + w \frac{\partial}{\partial z}$$

$$D^\phi \phi = m^2 \frac{\partial}{\partial \lambda} \left(\mu_\phi \frac{\partial \phi}{\partial \lambda} \right) + mn \frac{\partial}{\partial \theta} \left(\mu_\phi \frac{n}{m} \frac{\partial \phi}{\partial \theta} \right)$$

t	—	time
λ	—	longitude
θ	—	latitude
u, v, w	—	zonal, meridional and vertical velocity components
T, S, ρ , P	—	temperature, salinity, deviations of density from the mean density $\rho_0 = 1.2 \text{g/cm}^3$ and pressure
f (T, S, P)	—	non linear functional dependence defined by the equation of state
$\nu = \nu_u = \nu_v, \nu_T, \nu_S$	—	coefficients of vertical turbulent viscosity and diffusion of heat and salt.
$\mu = \mu_u = \mu_v, \mu_T, \mu_S$	—	coefficients of horizontal turbulent viscosity and diffusion of heat and salt.

l	–	$2\omega \sin\theta$, the coriolis parameter where ω is the earth's rotation speed
$M = n/\cos\theta$; $n = 1/a$; `		where 'a' is the radius of the Earth
$g = 981 \text{ gm/cm}^2$	–	acceleration due to gravity
τ^r, τ^θ	–	components of the tangential wind stress at the sea surface
τ_b^r, τ_b^θ	–	components of bottom stress
Q_T, Q_S	–	heat and salt flux or values of temperature and salinity depending on the values of the coefficients $\alpha_T, \beta_T, \alpha_S$ and β_S .

2.2.2 Model equations, boundary and initial conditions in σ - coordinates:

Equations (2.1)-(2.7) are transferred to the topographically conformal sigma coordinate using the transformation relation,

$$\sigma = Z/H \quad \text{_____} \quad (2.14)$$

where H is the total ocean depth and Z is the physical vertical coordinate. Its vertical coordinate, σ , is a dimensionless variable whose value ranges from zero at the surface to 1 at the bottom.

The final model equations in the sigma coordinate system are written as follows:

$$\frac{H}{m} \frac{du}{dt} - \frac{H}{m} (l + m \sin \theta u)v = -\frac{1}{\rho_0} \left(H \frac{\partial P}{\partial \lambda} - \sigma \frac{\partial H}{\partial \lambda} \frac{\partial P}{\partial \sigma} \right) + \tilde{D}'' u + \frac{\partial}{\partial \sigma} \left(\frac{v}{mH} \frac{\partial u}{\partial \sigma} \right)$$

_____ (2.15)

$$\frac{H}{m} \frac{dv}{dt} + \frac{H}{m} (l + m \sin \theta u) u = -\frac{n}{\rho_0 m} \left(H \frac{\partial P}{\partial \theta} - \sigma \frac{\partial H}{\partial \theta} \frac{\partial P}{\partial \sigma} \right) + \tilde{D}^u v + \frac{\partial}{\partial \sigma} \left(\frac{v}{mH} \frac{\partial v}{\partial \sigma} \right) \quad (2.16)$$

$$\frac{\partial P}{\partial \sigma} = gH\rho \quad (2.17)$$

$$\frac{duH}{d\lambda} + \frac{\partial}{\partial \theta} \left(\frac{mvH}{m} \right) + \frac{\partial}{\partial \sigma} \left(\frac{\tilde{w}}{m} \right) = 0 \quad (2.18)$$

$$\frac{H}{m} \frac{dT}{dt} = \tilde{D}^T T + \frac{\partial}{\partial \sigma} \left(\frac{v_r}{mH} \frac{\partial T}{\partial \sigma} \right) \quad (2.19)$$

$$\frac{H}{m} \frac{dS}{dt} = \tilde{D}^S S + \frac{\partial}{\partial \sigma} \left(\frac{v_s}{mH} \frac{\partial S}{\partial \sigma} \right) \quad (2.20)$$

$$\rho = f(T, S, P) - \rho_0 \quad (2.21)$$

where ;

$$\frac{H}{m} \frac{d}{dt} = \frac{H}{m} \frac{\partial}{\partial t} + uH \frac{\partial}{\partial \lambda} + \frac{n}{m} vH \frac{\partial}{\partial \theta} + \frac{1}{m} \tilde{w} \frac{\partial}{\partial \sigma}$$

$$\tilde{D}^\phi = m \frac{\partial}{\partial \lambda} \left[\mu_\phi \left(H \frac{\partial \phi}{\partial \lambda} - \sigma \frac{\partial H}{\partial \lambda} \frac{\partial \phi}{\partial \sigma} \right) \right] - m \frac{\partial}{\partial \sigma} \left[\frac{\mu_\phi}{H} \sigma \frac{\partial H}{\partial \lambda} \left(H \frac{\partial \phi}{\partial \lambda} - \sigma \frac{\partial H}{\partial \lambda} \frac{\partial \phi}{\partial \sigma} \right) \right]$$

$$+ n \frac{\partial}{\partial \theta} \left[\mu_\phi \frac{n}{m} \left(H \frac{\partial \phi}{\partial \theta} - \sigma \frac{\partial H}{\partial \theta} \frac{\partial \phi}{\partial \sigma} \right) \right] - n \frac{\partial}{\partial \sigma} \left[\frac{\mu_\phi}{H} \frac{n}{m} \sigma \frac{\partial H}{\partial \theta} \left(H \frac{\partial \phi}{\partial \theta} - \sigma \frac{\partial H}{\partial \theta} \frac{\partial \phi}{\partial \sigma} \right) \right]$$

The vertical velocity \tilde{w} can be written as

$$\tilde{w} = w - \sigma \left(mu \frac{\partial H}{\partial \lambda} + nv \frac{\partial H}{\partial \theta} \right)$$

Boundary conditions for the system (2.15) – (2.21) are as follows ;

At the sea surface ($\sigma = 0$)

$$\tilde{w} = 0 \quad ; \quad \frac{v}{H} \frac{\partial u}{\partial \sigma} = -\frac{\tau^\lambda}{\rho_0} \quad ; \quad \frac{v}{H} \frac{\partial v}{\partial \sigma} = -\frac{\tau^\theta}{\rho_0} \quad \text{_____ (2.22)}$$

$$\beta_T \frac{v_T}{H} \frac{\partial T}{\partial \sigma} + \alpha_T T = Q_T \quad ; \quad \beta_S \frac{v_S}{H} \frac{\partial S}{\partial \sigma} + \alpha_S S = Q_S$$

At the bottom ($\sigma = 1$) ;

$$\tilde{w} = 0 \quad ; \quad \frac{v}{H} \frac{\partial u}{\partial \sigma} = -\frac{\tau_b^\lambda}{\rho_0} \quad ; \quad \frac{v}{H} \frac{\partial v}{\partial \sigma} = -\frac{\tau_b^\theta}{\rho_0} \quad \text{_____ (2.23)}$$

$$\frac{v_T}{H} \frac{\partial T}{\partial N_\phi} = 0 \quad ; \quad \frac{v_S}{H} \frac{\partial S}{\partial N_\phi} = 0$$

Along the rigid lateral boundaries of the model, normal flux boundary condition of zero velocity are prescribed.

ie., $u = v = 0$; for temperature and salinity, the following boundary conditions were applied

$$\frac{\partial T}{\partial N_\phi} = 0 \quad ; \quad \frac{\partial S}{\partial N_\phi} = 0 \quad \text{_____ (2.24)}$$

The initial conditions are ;

$$u = u_0 \quad ; \quad v = v_0 \quad ; \quad T = T_0 \quad ; \quad S = S_0 \quad \text{_____ (2.25)}$$

The model also has an equation for stream function of the following form (Zalesny e al ., 2000)

$$\frac{\partial}{\partial t} A\phi + J\left(\phi, \frac{l}{H}\right) + RA\phi = \frac{gn}{\rho_0} J\left(H, \int_0^1 \sigma \rho \partial \sigma\right) \quad \text{_____ (2.26)}$$

The operator A and Jacobian, J, have the following form;

$$A\phi = \frac{\partial}{\partial \lambda} \frac{m}{nH} \frac{\partial \phi}{\partial \lambda} + \frac{\partial}{\partial \theta} \frac{n}{mH} \frac{\partial \phi}{\partial \theta}$$

$$J(\phi, z) = \frac{\partial \phi}{\partial \lambda} \frac{\partial z}{\partial \theta} - \frac{\partial \phi}{\partial \theta} \frac{\partial z}{\partial \lambda} \quad \text{_____ (2.27)}$$

In eq. (2.26), R is the relaxation coefficient. Equations (2.15)-(2.21) together with the equation for the stream function (2.26) form the full set of model equations in sigma coordinate system that are used in the present model study. While the system (2.15)-(2.21) were solved by the method of splitting, the equation for stream function (2.26) was solved by the method of successive overrelaxation.

2.2.3 Features and parameters of the σ - coordinate model

The main peculiarity of our model, which distinguishes it from other well known ocean model such as GFDL Modular Ocean Model (Pacanowski and Griffies, 1999) based on the pioneering works by Bryan (1969), Semtner (1974) and Cox (1984), or the σ -coordinate Princeton Ocean Model (Blumberg and Mellor, 1987), is that the numerical implementation of the model uses splitting method with respect to physical process and spatial coordinates. The splitting method allows one to implement efficient implicit time-integration schemes for the transport and diffusion equations; The Crank- Nicholson approximation was used for transport processes and implicit scheme was used for diffusion and second order viscosity terms. The Coriolis terms were also approximated implicitly. Implicit methods applied to time integration make it possible to use time steps several time greater than those in general circulation models based on explicit numerical scheme. This substantially reduces the CPU time required by the model which is especially important when long-term integration periods are needed for achieving quasi-equilibrium state. This version of the model uses spatial approximations on a staggered B-grid (Mesinger and Arakawa, 1976).

In this version of the model, it is possible to parameterize the horizontal viscosity by second and fourth order operations with explicit time stepping. The explicit approximation does not significantly restrict the choice of the time step which is associated with the use of different meridional and zonal viscosity values, and stability occurs in a wide range of time steps when different meridional and zonal viscosity terms are used .

In the numerical experiments, the horizontal diffusivities for temperature and salinity were set identical. The meridional diffusivity A_θ was specified as a function of latitude and depth by the following relation;

$$A_{\theta} = A[0.2 + 0.8\exp(\frac{\sigma H}{D})] * [0.5 + 0.5(\cos\theta)^{10}] \quad \text{_____ (2.28)}$$

Where $A = 1.5 \times 10^3 \text{ m}^2 \text{ s}^{-1}$ is the maximum of A_{θ} attained at the ocean surface and equator simultaneously and $D = 300$ is the e-folding parameter and θ is the latitude. The exponent 10 for the cosine of latitude was chosen so that the diffusivity is maximum at the equator, gradually decreases to a half at 30° and virtually constant as the latitude increases further. The diffusivities are specified in such a manner so that the velocity and vorticity reach their maximum values in the equatorial regions. The zonal diffusivity A_{λ} was defined as :

$$A_{\lambda} = \left(\frac{A_{\theta} \Delta\lambda \cos\theta}{\Delta\theta} \right) \quad \text{_____ (2.29)}$$

Where $\Delta\lambda$ and $\Delta\theta$ are the longitudinal and latitudinal model space steps, expressed in degrees or radians.

The meridional and zonal second order lateral viscosities were specified in a similar manner as diffusivity, but without decreasing with depth. In this version of the model, the second order viscosity plays an auxiliary role. The meridional viscosity is specified by a small background value not exceeding $2 \text{ m}^2/\text{s}$. The numerical noise in the horizontal velocity field is suppressed primarily by fourth order viscosity with a constant meridional viscosity $B_{\theta} = 5.0 \times 10^{14} \text{ m}^4 \text{ s}^{-1}$ and variable zonal viscosity

$$B_{\lambda} = B_0 \left(\frac{\Delta\lambda \cos\theta}{\Delta\theta} \right)^4$$

The vertical viscosity and diffusivities were parameterized according to Pacanowski and Philander parameterization (1981). The viscosity ranged

from 15×10^{-4} to $90 \times 10^{-4} \text{ m}^2\text{s}^{-1}$ and diffusivity varied from the background value 0.15×10^{-4} to $50.15 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ as Ri decreased from large values to zero. In case of unstable stratification when $Ri < 0$, the vertical diffusivity was increased to $500 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ and the vertical viscosity was set equal to $90 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$.

2.2.4 Geometry, resolution and topography

The model domain extends from 36°S to 29.5°N in the meridional direction and from 22°E to 142°E in the zonal directions. The land boundaries used in the present study are shown in Fig.1. The model domain includes Red Sea, Persian Gulf and islands of Madagascar and Sri Lanka. The model domain also includes two liquid boundaries. The large southern liquid boundary is placed at 36°S and the eastern small boundary is placed at the vicinity of Moluccas islands. The model has a moderate resolution of $1^\circ \times 1/2^\circ$ in the longitude and latitude directions respectively and a vertical resolution of 33 unequal sigma levels. In a water column of 3500 m depth, there are 9 levels in the upper 150m of the water column and the model levels are selected in such a way that the mixed layer process, equatorial and near equatorial current are well resolved.

The salinity structure in the north Indian Ocean is considerably affected by the intrusion of Red Sea and Persian gulf waters and the inclusion of these two marginal seas into the model would considerably improve the simulated field of salinity. The runoff of major rivers that influence the salinity distribution must also be taken into account while formulating numerical model of large scale circulation. In this version of the model, the runoff of rivers Brahmaputra, Ganges and Indus contributing considerably to the salinity distribution are taken into account. The procedure for incorporating river runoff in the present model will be discussed in chapter 4.

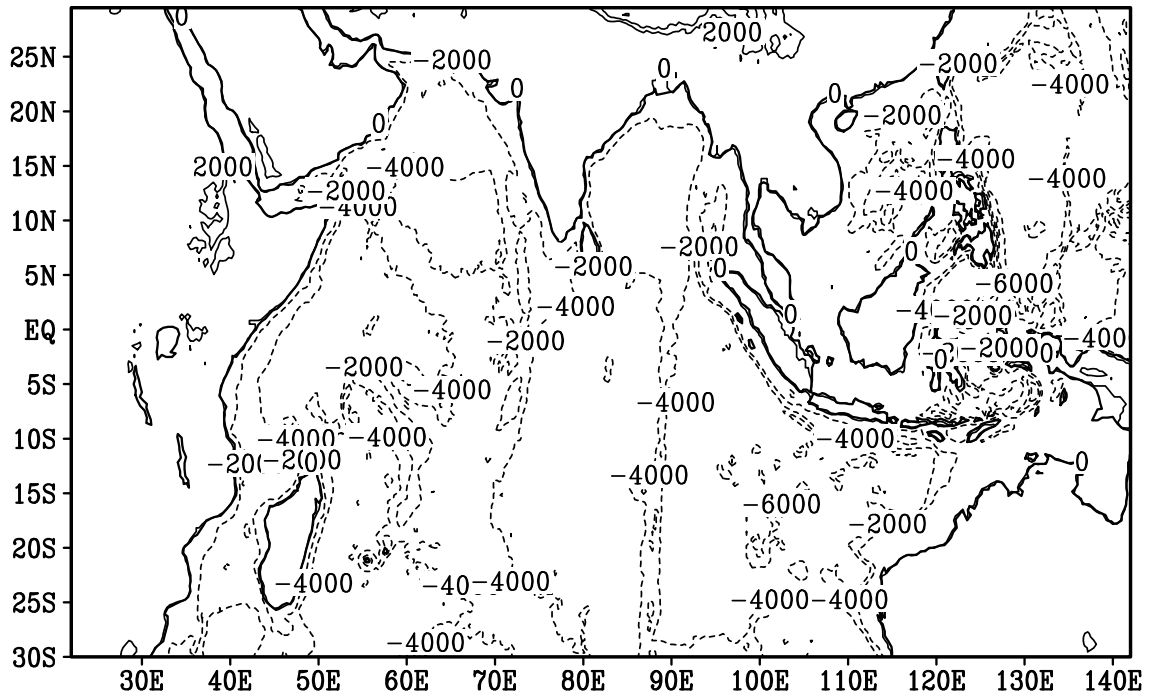


Fig.1 Bottom topography of the model area driven from ETOPO5 data set; depths are in meters

The model bottom topography is derived from the 5' resolution ETOPO5 data set supplied by the National Geophysical Data Centre, USA. A non-linear filter was used to mitigate topographic gradients that would cause spurious alongshore currents in sigma coordinate model.

2.3 Surface forcing parameters

The large scale circulation in the ocean is a complicated function of the internal density structure of the water mass, radiative and turbulent heat fluxes at the surface, the wind stress and buoyancy fluxes imposed at the sea surface by the overlying atmosphere. In the present work, Levitus data sets on temperature and salinity (Levitus, 1994) were used to initialise the models. Monthly mean meteorological fields derived from National Center for Environmental Prediction/National Center for Climate Research (NCEP/NCAR) reanalysis data (Kalnay et al. 1996) were used to calculate wind stress, heat and fresh water fluxes. In the sigma coordinate ocean model, wind stress, heat and fresh water fluxes are the main surface forcings and these forcing parameters, derived from NCEP/NCAR reanalysis data and averaged over the period 1982 to 1994, were bilinearly interpolated to the model grid points. The wind stress components were interpolated to the Arakawa B-grid, where horizontal velocities were computed. Heat and fresh water fluxes were bilinearly interpolated to the grid points where temperature and salinity values were computed. Details of computations of the above mentioned fluxes from NCEP/NCAR reanalysis data sets are presented in the under mentioned paragraphs.

2.3.1 Wind Field;

The NCEP reanalysis project used the state of art analysis system, GDAS (Global Data Assimilating System) to hind cast meteorological fields from 1957 onwards. The project involves the recovery of land surface , ship, radiosonde, aircraft, satellite, and other data, quality controlling and

assimilating these data with GDAS that is kept unchanged over the reanalysis period. This eliminates artificial jumps. In addition to using a state of art analysis scheme, the reanalysis also used delayed observations, increasing the reliability of the surface wind products. The data were then bileanearly interpolated to the Indian Ocean circulation model grid points.

The wind stress is computed using the bulk aerodynamic formula

$$\tau = \rho_a C_D |v|v \quad \text{_____} \quad (2.30)$$

Where $\rho_a = 1.2 \text{ Kg m}^{-3}$ is the density of air, $C_D = 1.03 \times 10^{-3}$ is the drag coefficient : The wind stress computed from the reanalysis for the four cardinal months of the year, namely January, July , April and October, representing the four season of winter, summer, spring and Autumn for the model area are presented in the next chapter.

2.3.2 Net heat flux

The heat-flux components for forcing the model were derived from the NCEP reanalysis data. The net heat flux is calculated using the formula .

$$Q = Q_b + 0.6 Q_s + Q_{\text{sen}} + Q_{\text{lhf}} + \Gamma_\theta(\theta_L - \theta_1) \quad \text{_____} \quad (2.31)$$

Q_b is the longwage radiation

Q_s is the short wage radiation

Q_{sen} is the sensibtle heat flux

Q_{lhf} is the latent heat flux

θ_L is the Levitus surface temperature

θ_1 is the temperature at the first model level

Γ_θ is the restoring coefficient.

The shortwave radiation undergoes absorption and scattering while passing through the sea water. The total heat flux is mostly absorbed in the first level (10m), but, since the near surface resolution is very high, account is taken for the penetration of solar radiation into the sea. In the first 35cm of the water column, 60% of the total shortwave radiation incident on the sea surface is absorbed and the rest undergoes exponential decay with a parameter of 20m while penetrating through the water column. The coefficient 0.6 of Q_s is introduced because it is this portion of the shortwave radiation flux that is added to the total heat flux at the ocean surface (Diansky, N. A. et al., 2002) .

The last term in the equation is the restoring component of the net heat flux and it represent the negative feed back that drives the model SST to the observed climatic cycles. The values chosen for Γ_θ can be interpreted as the relaxation of the model temperature averaged over the upper 50m layer to the observed values with a 12 day time scale. This relatively strong relaxation forces the model to closely follow with the observations. The analysis of NCEP data fields may have some deviation from the real climate. so, the restoring terms are incorporated in the forcing fields to prevent the deviation of the model climate from the observed one.

2.3.3 Fresh water flux

The mass flux is determined by precipitation minus evaporation and is implemented as salt fluxes.

$$F = P - E - \Gamma_F (S_L - S_1) \quad \text{_____} \quad (2.32)$$

Where ;

P is the precipitation

E is the evaporation and the last term are restoring terms

S_1 is the model simulated upper layer salinity.

S_L is the Levitus upper Layer salinity

Γ_F is the restoring coefficient.

The freshwater flux from the Ganges and Brahmaputra rivers into the Bay of Bengal and the fluxes from Indus into Arabian Se are also included in the model simulation.

2.3.4 Time interpolation of surface forcing parameters

In prognostic models of ocean circulation, all surface forcing parameters such as wind stress components, sea surface temperature and sea surface salinity are to be interpolated in time at each time step of model intergration. In the present work, we used a procedure for the time interpolation of surface forcing fields which considered the actual distribution of days in the months of the year including leap years and also preserved the monthly means of the interpolated fields. The formulae used are :

$$f(t_m) = f_{m-1} \left(\frac{1}{2} - \frac{3}{2} \frac{t_m}{T_m} \right) + f_m \left(\frac{1}{2} + 2 \frac{t_m}{T_m} \right) + f_{m+1} \left(-\frac{t_m}{2T_m} \right),$$

$$0 < t_m < \frac{T_m}{2} \quad \text{_____ (2.33)}$$

$$f(t_m) = f_{m-1} \left(-\frac{1}{2} + \frac{1}{2} \frac{t_m}{T_m} \right) + f_m \left(\frac{5}{2} - 2 \frac{t_m}{T_m} \right) + f_{m+1} \left(-1 + \frac{3t_m}{2T_m} \right),$$

$$\frac{T_m}{2} < t_m \leq T_m \quad \text{_____ (2.34)}$$

Here $f(t_m)$ is the interpolated field at time t_m measured from the beginning of the current month with index 'm' and duration T_m ; f_{m-1} , f_m and f_{m+1} are the monthly means of original field for the preceding, current and succeeding months. It can be seen that the relations (2.33) and (2.34) satisfies the relation:

$$\int_0^{T_m} f(t_m) dt_m = f_m$$

The above mentioned interpolation formulai allow one to reproduce monthly mean climatologies of SST and SSS to relatively very high accuracy.