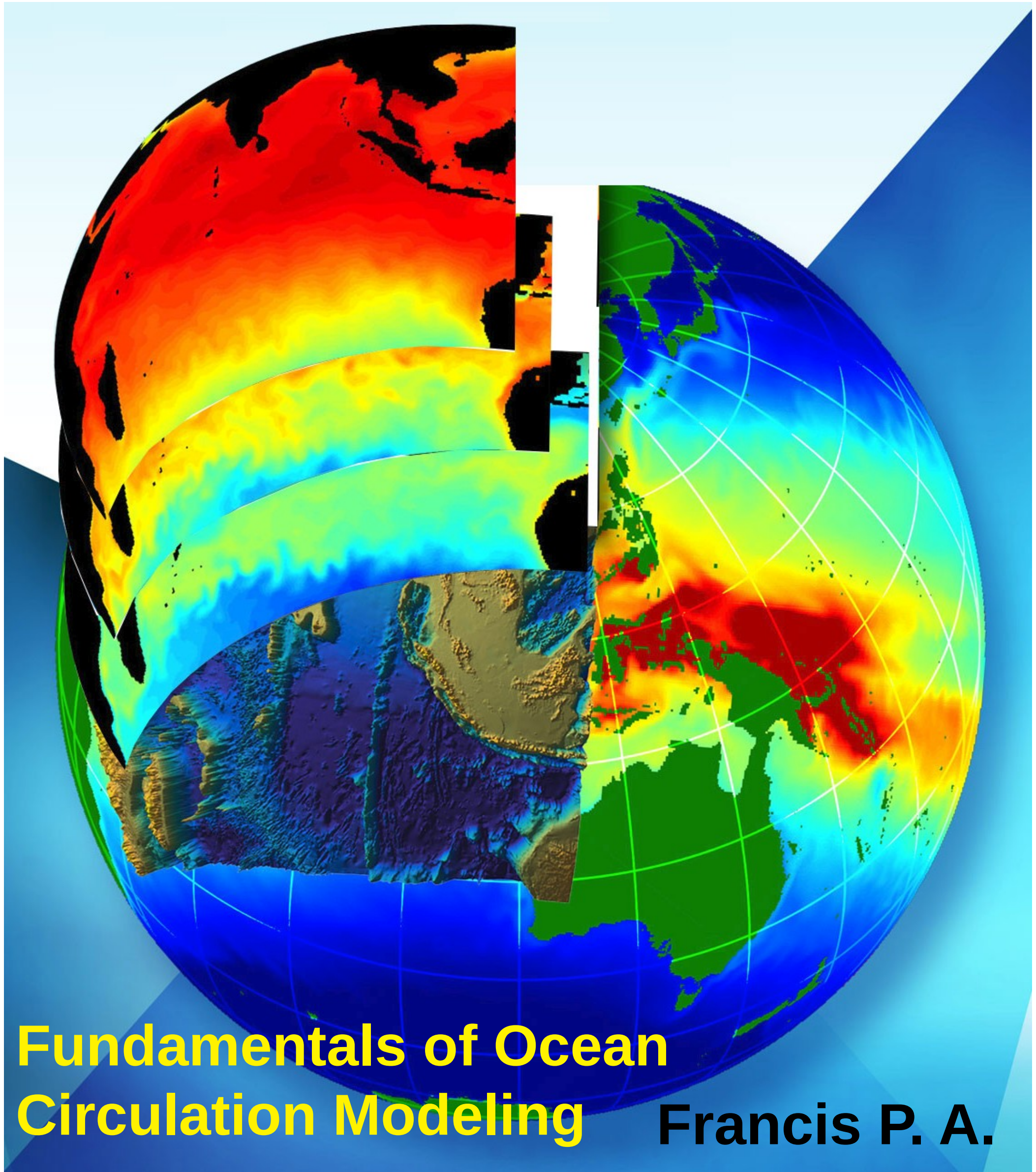
"Fundamentals of Ocean Circulation Modeling”



Francis P. A.

27th September 2021

(11:00-13:00 hrs IST; 0530:0730 hrs UTC)

International Training Centre for Operational Oceanography (ITCOocean)

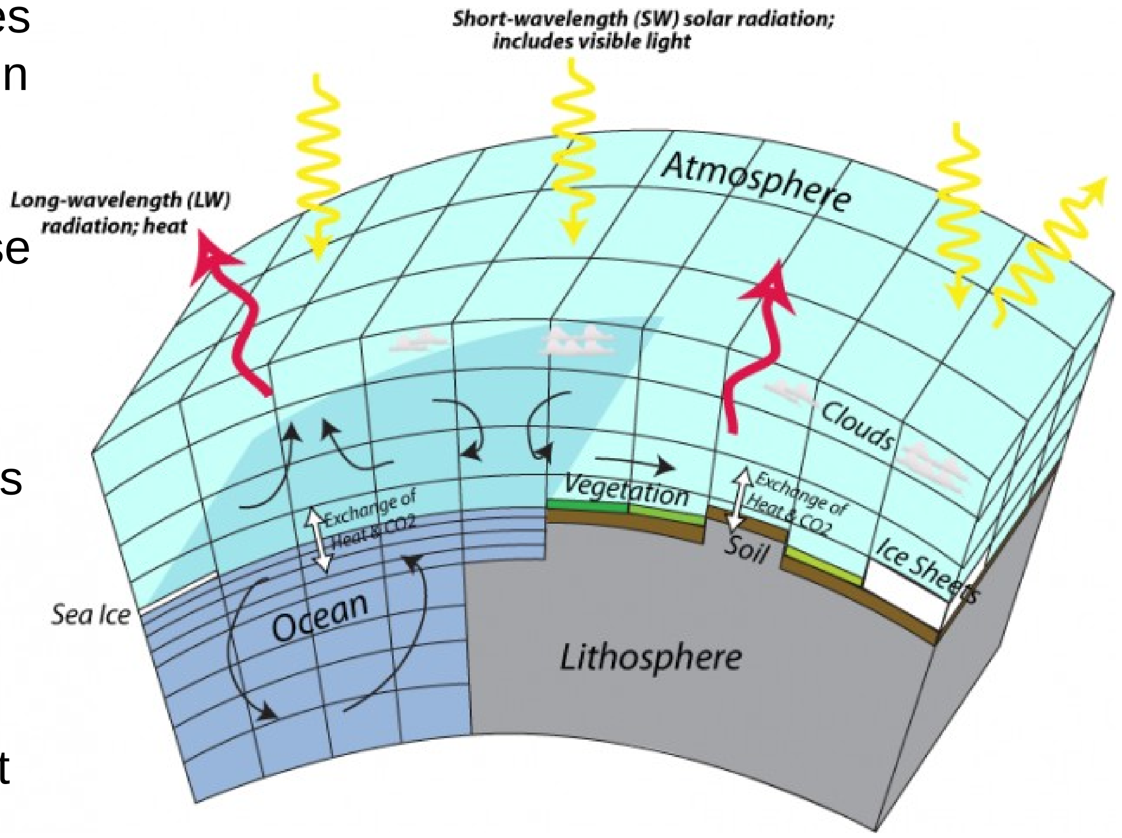
Indian National Centre for Ocean Information Services

Ministry of Earth Sciences, Govt. of India

Hyderabad-500090.

An introduction to climate models

- Large-scale ocean-atmosphere processes plays a crucial role in climate variability on a broad range of space and time scales
- Understanding the dynamics behind these processes are hence very important
- Ocean-atmosphere models are now become essential tools for climate studies
- Coupled and stand-alone models are integral part of the prediction systems.
- While the ocean and atmosphere interact each other at all time-scales, it is important to identify the most-critical processes so as to model the system efficiently and without loosing the essence of the basic climate system



Elements of ocean-atmoshere modeling system

Image Credit: David Bice © Penn State University

An introduction to circulation models

What are general circulation models?

- General circulation models are mathematical representation (most often in the form of partial differential equations) of the processes (physical/chemical/biological etc.) taking place in the earth system, which are discretised and solved numerically using powerful High Performance Computing Facilities, after making suitable assumptions and approximation and providing necessary initial and boundary conditions.

- Since many processes can not be resolved by the model equations due to many reasons, including the spacing of model grids, the effects of such processes are explicitly incorporated in the models (physical parameterisation)

Steps involved in General Circulation Modeling

- Observations
- Deriving Governing Equations
- Assumptions
- Approximations
- Co-ordinate Transformations
- Parameterisations
- Numerical discretisation
- Initial conditions
  - Data Assimilation
- Boundary conditions
- Coding
- Transforming into HPC environment
  - MPI/OpenMP env.
- Preprocessing
- Execution of model
  - Analysis/reanalysis/projections
- Post Processing
- Diagnosing/Analysis/Forecasts
- Interpretation/Products

Momentum Equations

$$\frac{du}{dt} - fv = -\frac{1}{\rho}\left(\frac{\partial p}{\partial x}\right) + Fr_x$$

$$\frac{dv}{dt} + fu = -\frac{1}{\rho}\left(\frac{\partial p}{\partial y}\right) + Fr_y$$

$$\frac{dw}{dt} = -\frac{1}{\rho}\left(\frac{\partial p}{\partial z}\right) - g + Fr_z$$

where

$$\frac{d}{dt} = u\frac{\partial}{\partial x} + \mathrm{v}\frac{\partial}{\partial y} + w\frac{\partial}{\partial z}$$

Friction (drag)

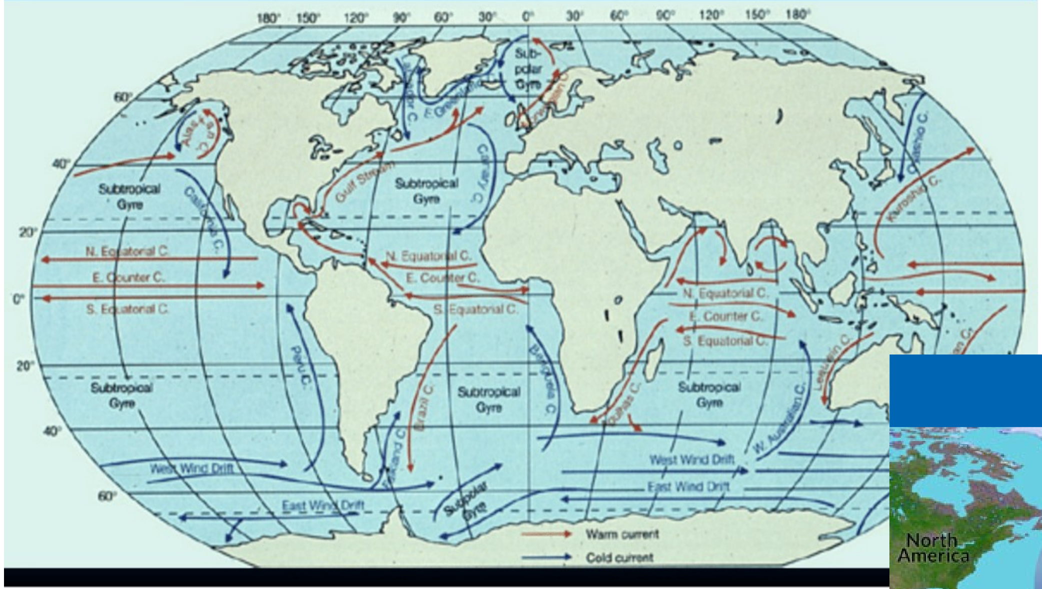
Surface drag is proportional to

- square of the wind speed
- Area of the fetch

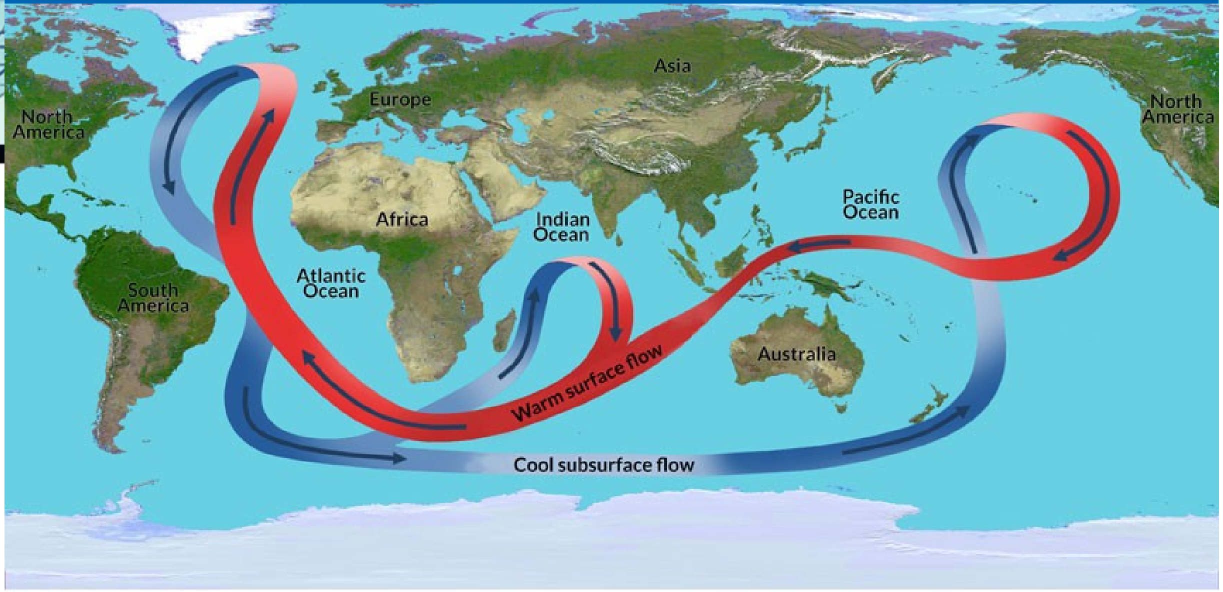
Pressure gradient force

Coriolis force

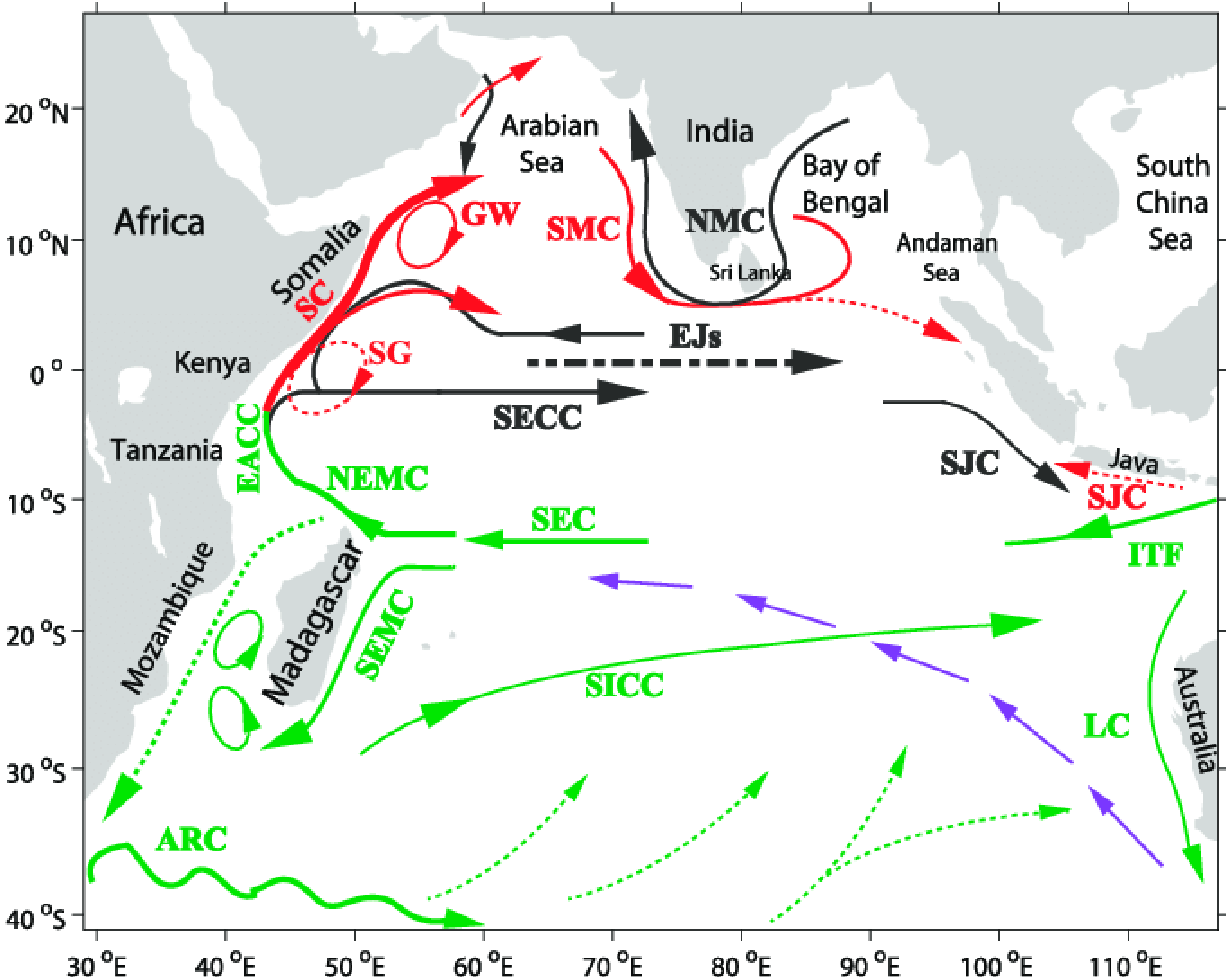
Wind-Driven Ocean Circulation

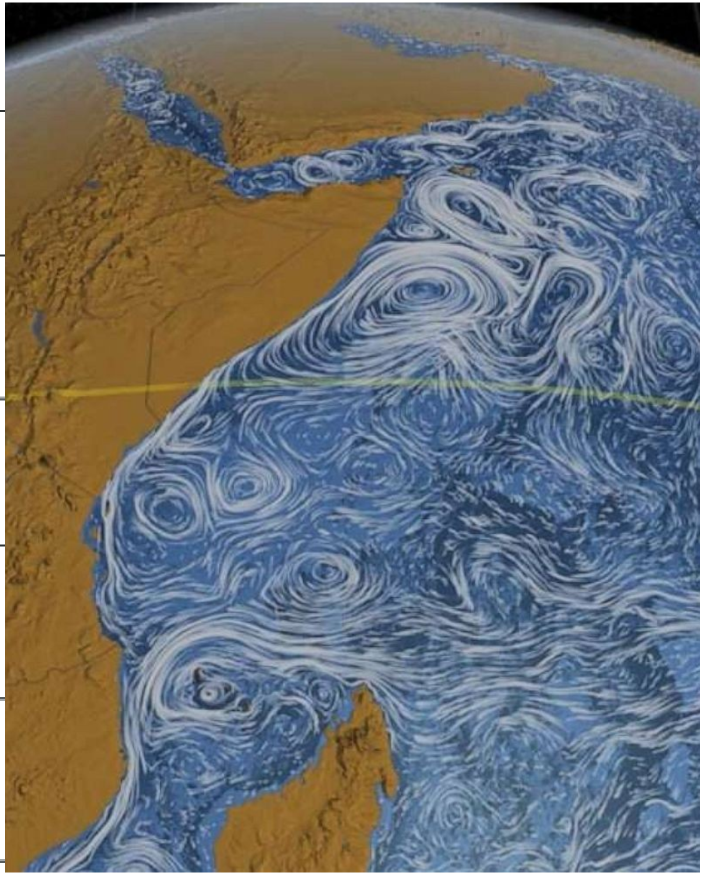


Thermo-haline Circulation

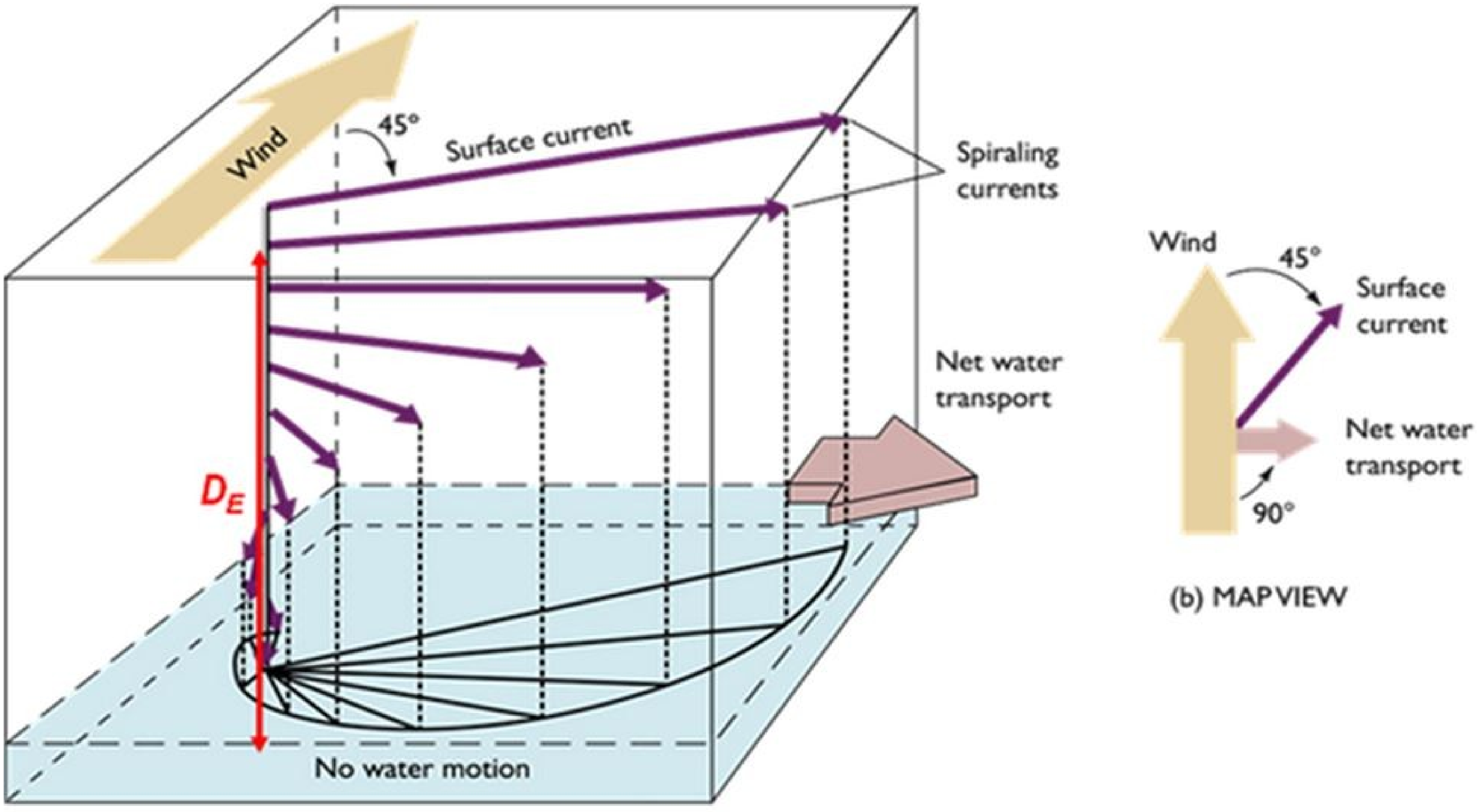


Indian Ocean Circulation





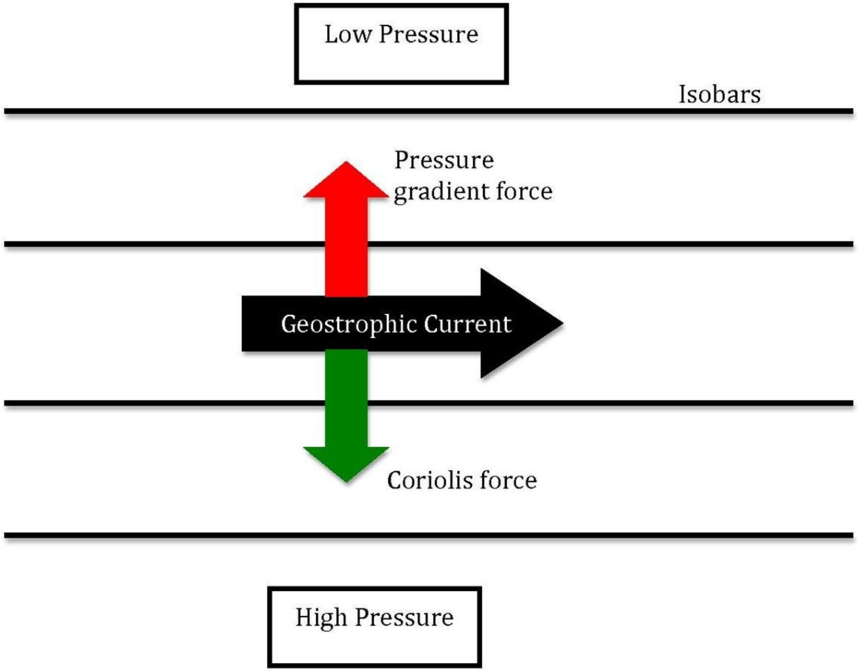
Ekman Drift



(a) EKMAN SPIRAL IN THE NORTHERN HEMISPHERE

(b) MAP VIEW

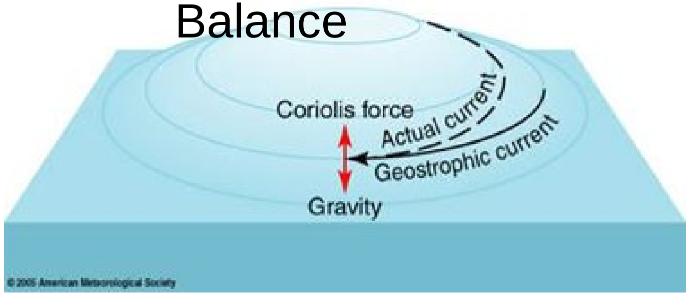
Geostrophic currents

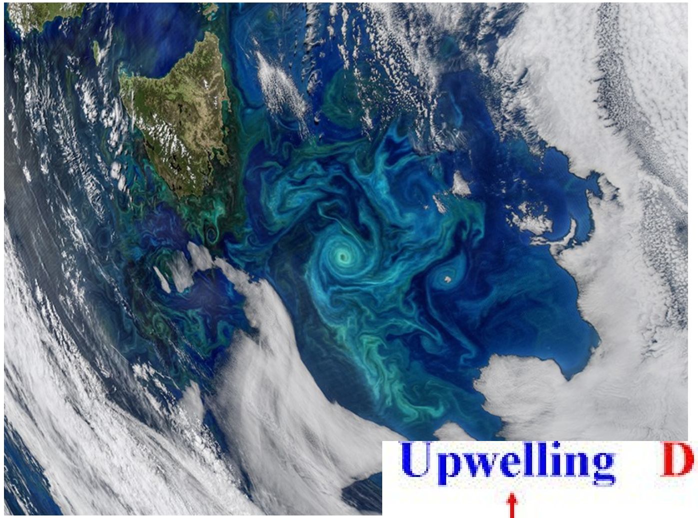


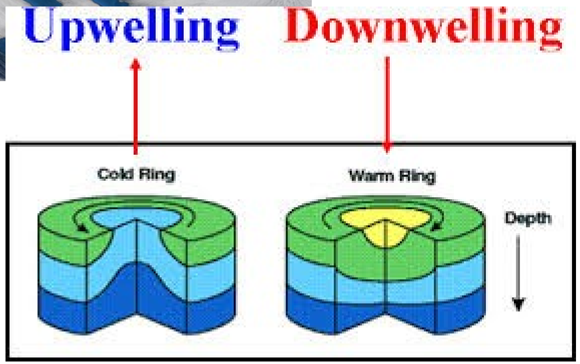
The equations can be written:

$$u = -\frac{1}{f\rho}\frac{\partial p}{\partial y}; \qquad v = \frac{1}{f\rho}\frac{\partial p}{\partial x}$$

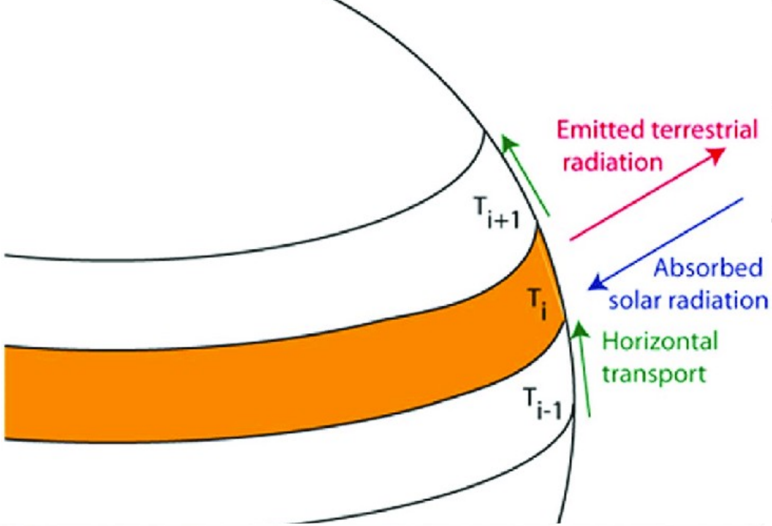
Geostrophic Balance

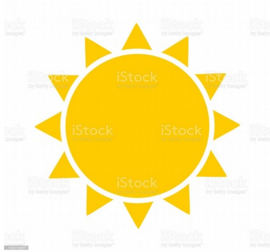


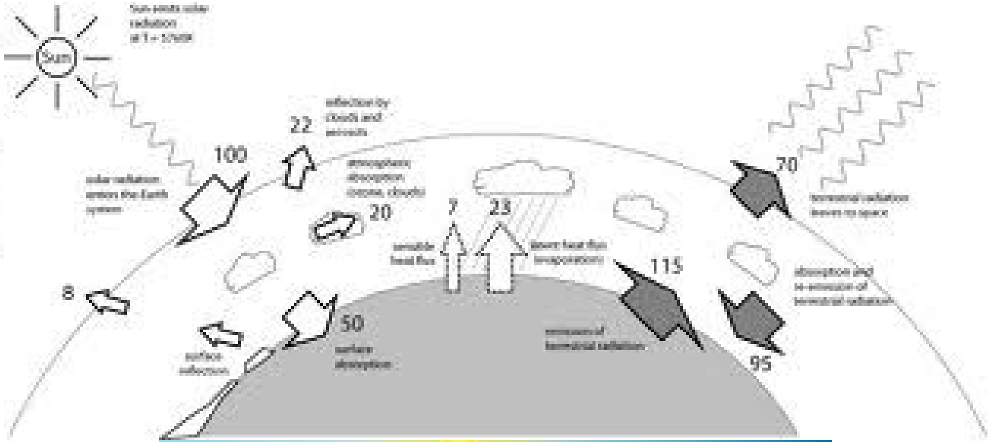


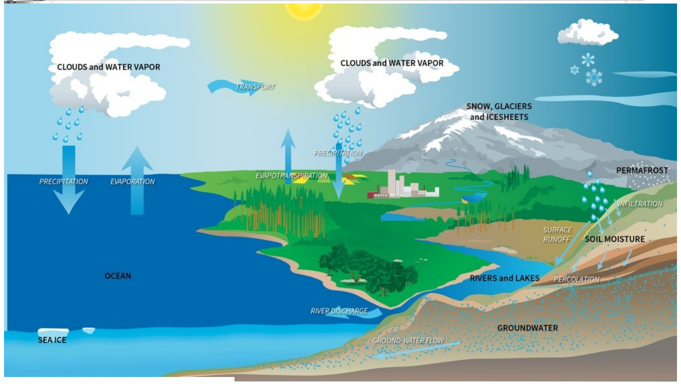


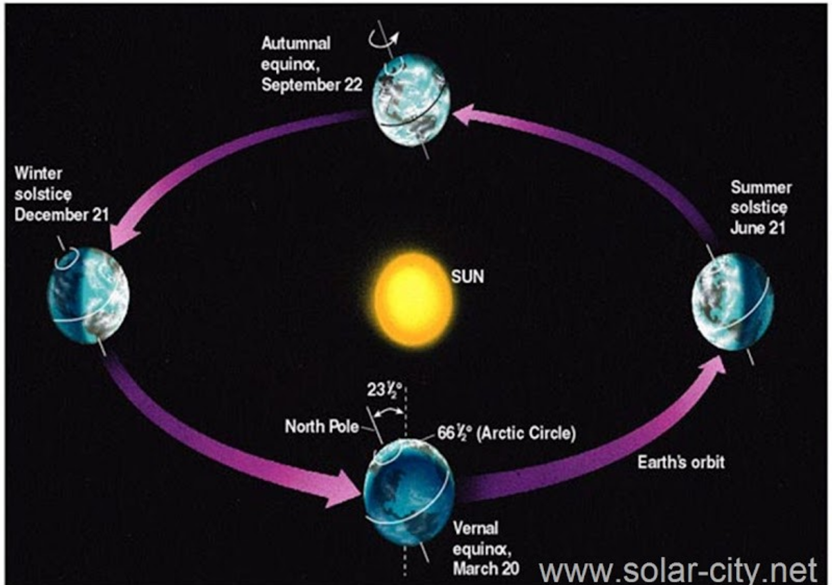
Radiation Budget: 1-D model

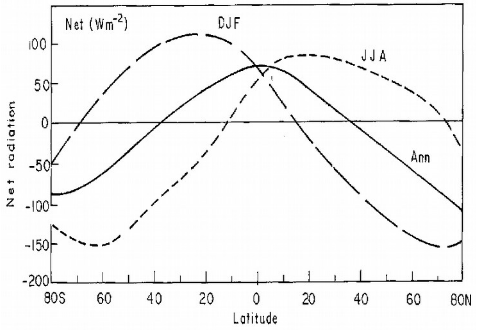




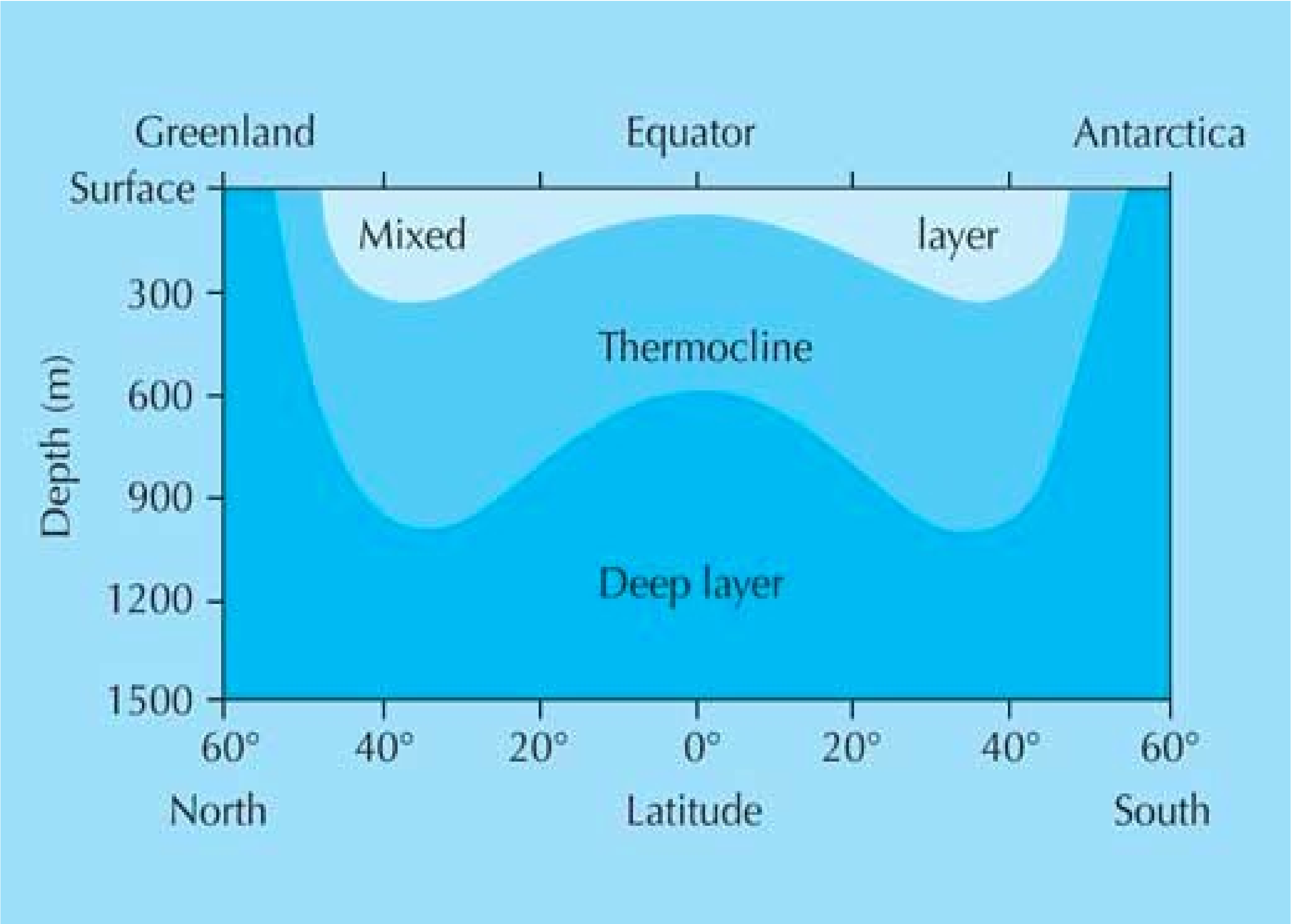








Latitudinal variation of ocean layers



- Skin
- Surface Mixed Layer
- Thermocline
- Interior Ocean

Primitive Equations: Hydrostatic, Incompressible, Boussinesq

Momentum

$$\begin{cases} \dfrac{\partial u}{\partial t} + \vec{v}\cdot\nabla u - fv = -\dfrac{\partial \phi}{\partial x} + \mathcal{F}_u + \mathcal{D}_u \\ \dfrac{\partial v}{\partial t} + \vec{v}\cdot\nabla v + fu = -\dfrac{\partial \phi}{\partial y} + \mathcal{F}_v + \mathcal{D}_v \end{cases}$$

Tracer

$$\begin{cases} \dfrac{\partial T}{\partial t} + \vec{v}\cdot\nabla T = \mathcal{F}_T + \mathcal{D}_T \\ \dfrac{\partial S}{\partial t} + \vec{v}\cdot\nabla S = \mathcal{F}_S + \mathcal{D}_S \end{cases}$$

$$\rho = \rho(T, S, P)$$

Hydrostatic

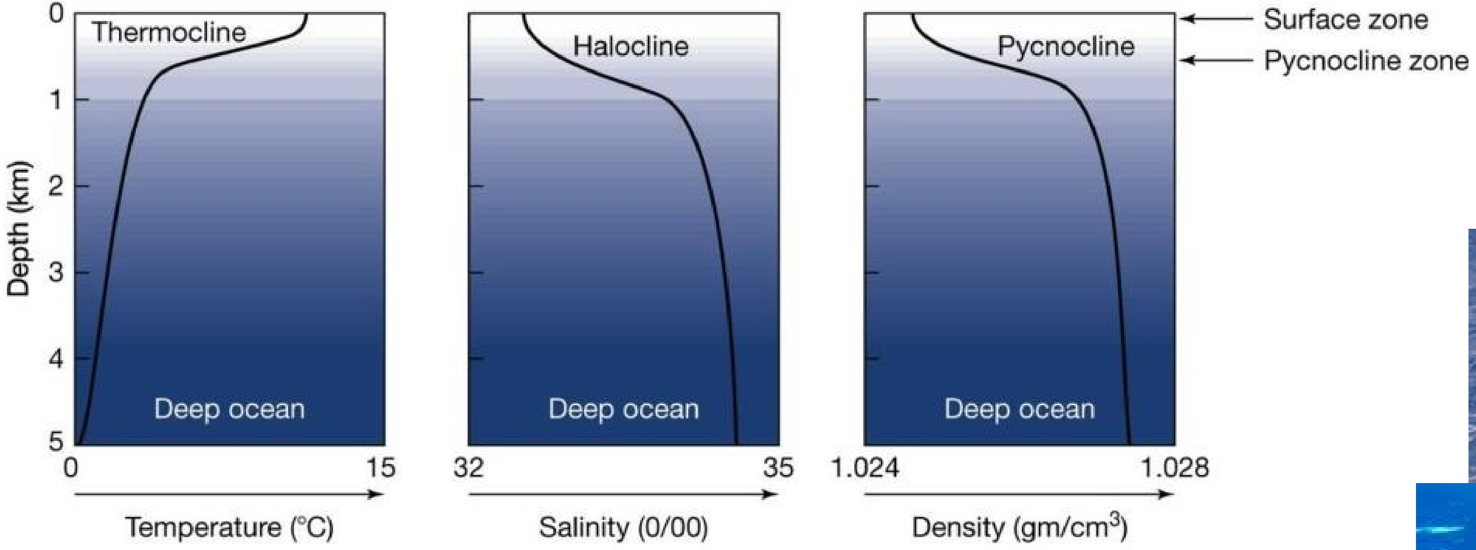
$$\frac{\partial \phi}{\partial z} = \frac{-\rho g}{\rho_o}$$

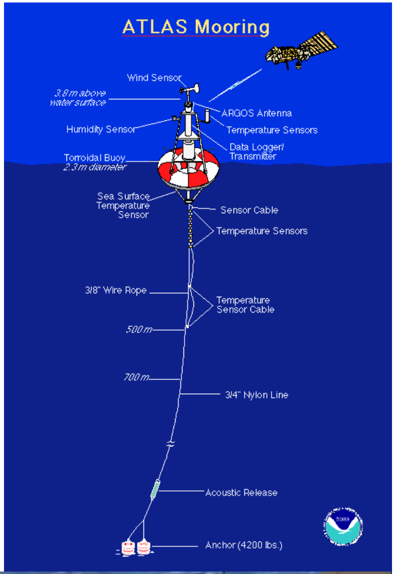
Continuity

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0.$$

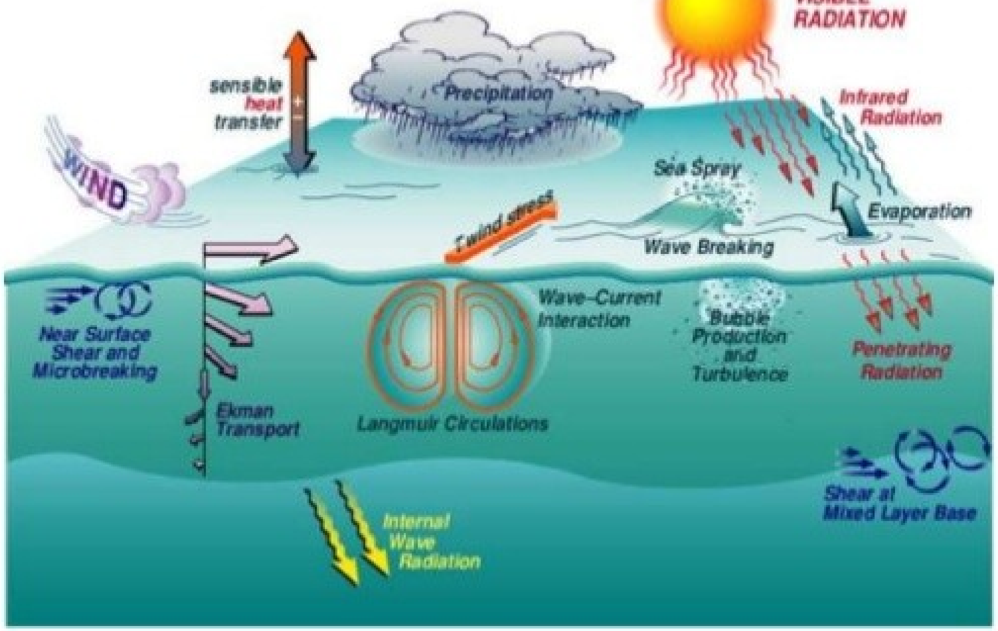
Similar transport equations for other tracers: passive or actives

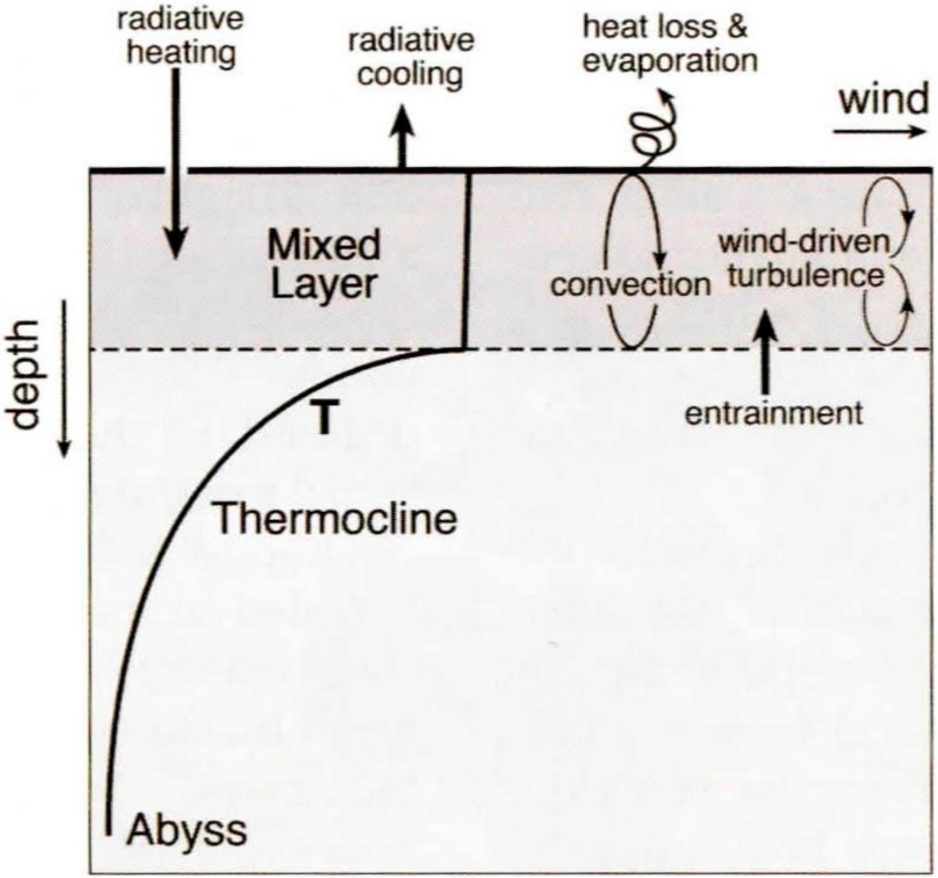
Vertical structure





Mixing Processes





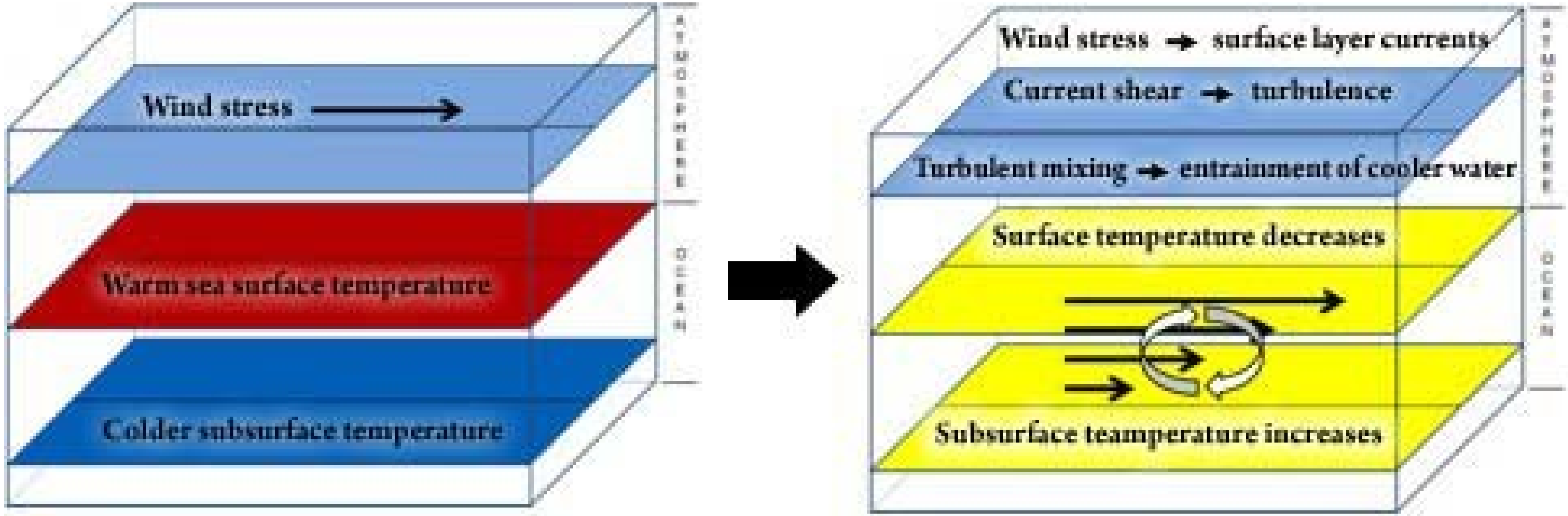
Surface mixed layer

1. Convection
2. Wind forcing
   - Wind driven Shear
   - Langmuir Circulation
   - Wave breaking
3. Precipitation
4. Sea Ice

$$\frac{\partial [S]}{\partial t} = \underbrace{\frac{(E-P)}{h}[S]}_{\text{SEF}} - \underbrace{[\nabla_H \cdot (uS, vS)]_{ml} + [\text{mixing}]_H}_{\text{Horizontal Advection}}$$

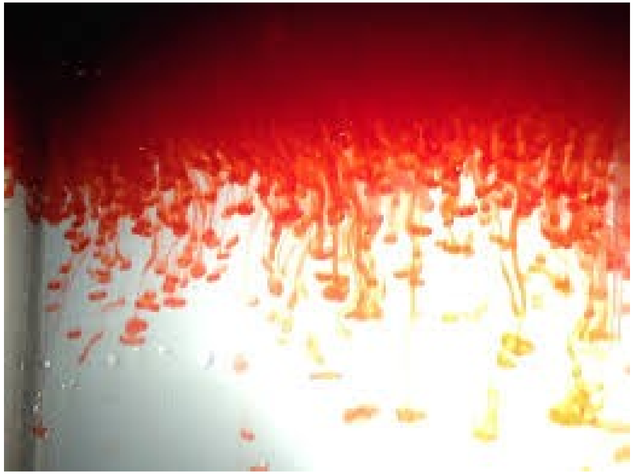
$$\underbrace{\underbrace{-\frac{1}{h}\Delta S \frac{\partial h}{\partial t} - [\nabla_H \cdot (uS, vS)]_{induct} - [\nabla_z (wS)]}_{\text{Vertical Entrainment}} + [\text{mixing}]_z,}_{\text{Subsurface Processes}}$$

Mixing Processes



Shear

Buoyancy



Net Effect?

Mixing Processes

Two dominant processes in the surface mixed layer

1. Vertical shear in the horizontal flow (du/dz)

2. Buoyancy (g/(rho)- Brunt Vaisala Frequency $N = \left(-\frac{g}{\rho_0} \frac{\partial \rho}{\partial z} \right)^{1/2},$

The ratio between these two terms are called Gradient Richardson Number

$$Ri = N^2 \Big/ \left(\frac{du}{dz} \right)^2 = -g \frac{d\rho}{dz} \Big/ \rho_0 \left(\frac{du}{dz} \right)^2,$$

Gradient Richardson Number and Bulk Richardson Number

Local and Non-local turbulance closure schemes

One of the popular mixing scheme based on bulk Richardson Number calculation is KPP scheme (Large et al. 1994) or LMD94 scheme.

Mixing Processes: KPP Scheme

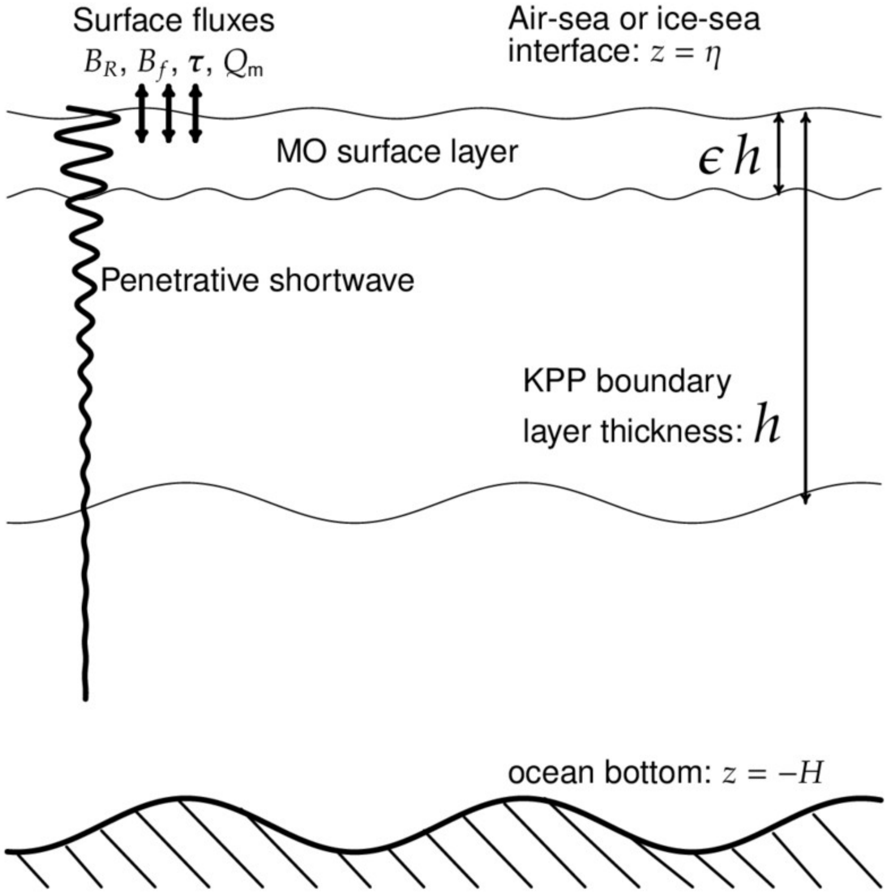
Mixing Processes taken into account by KPP Mixing Scheme

- A boundary layer can be defined as the layer of fluid in immediate vicinity of a bonding surface where the viscous effect on the flow is prominent.

- Hence for the analysis, Large et al. 1994 suggested to divide the domain of ocean in two parts:
  a) Oceanic Boundary Layer
  b) Interior Ocean

- Each of these domains have their own distinct governing equations, and thus have to be analyzed separately.

- Mixing is quantified using a parameter called Diffusivity (K_x).



MO: Monin Obukhov surface layer

Von-Roekel et al. (2017)

Boundary Layer :

- Boundary Layers are fundamentally turbulent and extend from the near surface to the Height of Surface Boundary Layer (HSBL), which determines the depth to which turbulent eddies can penetrate.

- These eddies are due to Wind Stress and Surface Buoyancy.

- The height of OBL is defined as the smallest depth where the bulk Richardson number equals a certain value called the critical Richardson number. This basically indicates that the boundary layer eddies should be able to penetrate to the depth 'h' before they become stable with respect to local buoyancy and velocity.

- In the boundary layer, the diffusivity (ν_x) is expressed as the product of a depth dependent turbulent velocity scale (w_x) , the height of OBL (h_{sbl}) and a non-dimensional shape function (G_x).
  - $\nu_x = h_{sbl} \, w_x(\sigma) \, G_x(\sigma)$

- As per tests conducted by O Brien (1970) in the Atmospheric Boundary layer, the shape function was assumed to be a cubic polynomial.
  - $G(\sigma) = a_o + a_1 \sigma + a_2 \sigma^2 + a_3 \sigma^3$

The coefficients (a_0, a_1, a_2, a_3) are found out by matching the boundary layer diffusivity at the base of the OBL to the diffusivity at the same depth of Interior Ocean.

Interior Ocean :

The following processes are taken into account for interior ocean mixing

- Interior Ocean Mixing due to Shear Mixing (using the Gradient Richardson number)
- Interior Ocean Mixing due to Internal Waves.
- Interior Ocean Mixing due to Double Diffusion
- (Diffusive convection & Salt Fingering)

Ri_g = Gradient Richardson Number

Ri_0 = Critical value of Ri_g

$$\nu_x(d) = \nu_x^s + \nu_x^d + \nu_x^w$$

$$\nu_x^s = \begin{cases} \nu_0 & Ri_g < 0, \\ \nu_0[1-(Ri_g/Ri_0)^2]^3 & 0 < Ri_g < Ri_0, \\ 0 & Ri_g > Ri_0. \end{cases}$$

where ν_0 is 5.0×10^{-3}, $Ri_0 = 0.7$.

$$\nu_s^d(R_\rho) = \begin{cases} 1 \times 10^{-4}[1-(\frac{(R_\rho-1)}{R_\rho^0-1})^2]^3 & \text{for } 1.0 < R_\rho < R_\rho^0 = 1.9, \\ 0 & \text{otherwise.} \end{cases}$$

$$\nu_\theta^d(R_\rho) = 0.7\nu_s^d$$

Internal wave generated mixing serves as the background mixing in the LMD scheme. It is specified as a constant for both scalars and momentum. Eddy diffusivity is estimated based on the data of Ledwell et al. (1993), while Peters et al. (1988) suggest eddy viscosity should be 7 to 10 times larger than diffusivity for gradient Richardson numbers below approximately 0.7. Therefore LMD use

$$\nu_m^w = 1.0 \times 10^{-4} m^2 s^{-1}$$

$$\nu_s^w = 1.0 \times 10^{-5} m^2 s^{-1}$$

Tracer & momentum

Double diffusion density ratio

$$R_\rho = \frac{\alpha \frac{\partial \bar{\theta}}{\partial z}}{\beta \frac{\partial \bar{S}}{\partial z}},$$

Primitive Equations: Hydrostatic, Incompressible, Boussinesq

Momentum

$$\begin{cases} \dfrac{\partial u}{\partial t} + \vec{v}\cdot\nabla u - fv = -\dfrac{\partial \phi}{\partial x} + \mathcal{F}_u + \mathcal{D}_u \\ \dfrac{\partial v}{\partial t} + \vec{v}\cdot\nabla v + fu = -\dfrac{\partial \phi}{\partial y} + \mathcal{F}_v + \mathcal{D}_v \end{cases}$$

Tracer

$$\begin{cases} \dfrac{\partial T}{\partial t} + \vec{v}\cdot\nabla T = \mathcal{F}_T + \mathcal{D}_T \\ \dfrac{\partial S}{\partial t} + \vec{v}\cdot\nabla S = \mathcal{F}_S + \mathcal{D}_S \end{cases}$$

$$\rho = \rho(T,S,P)$$

Hydrostatic

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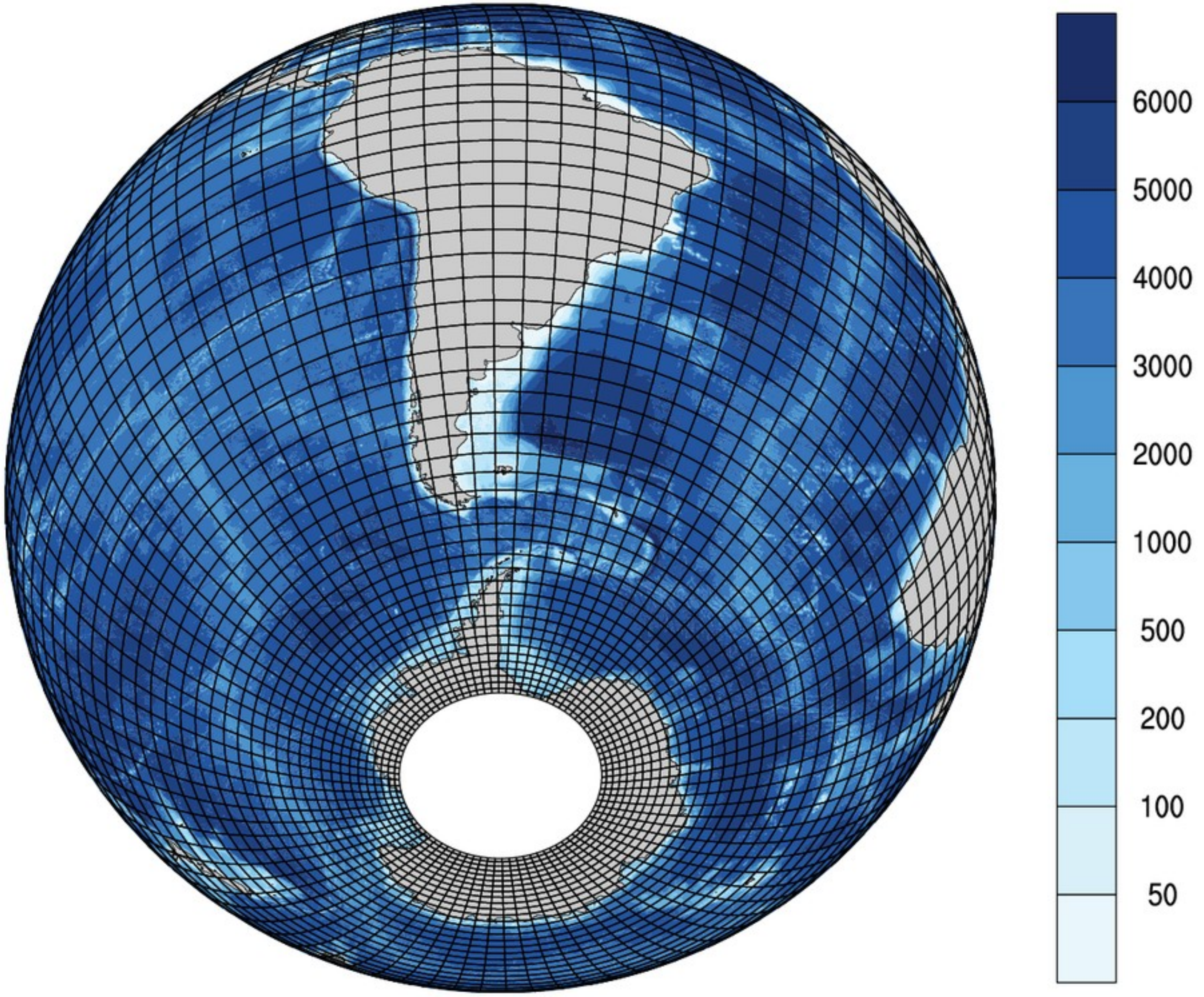
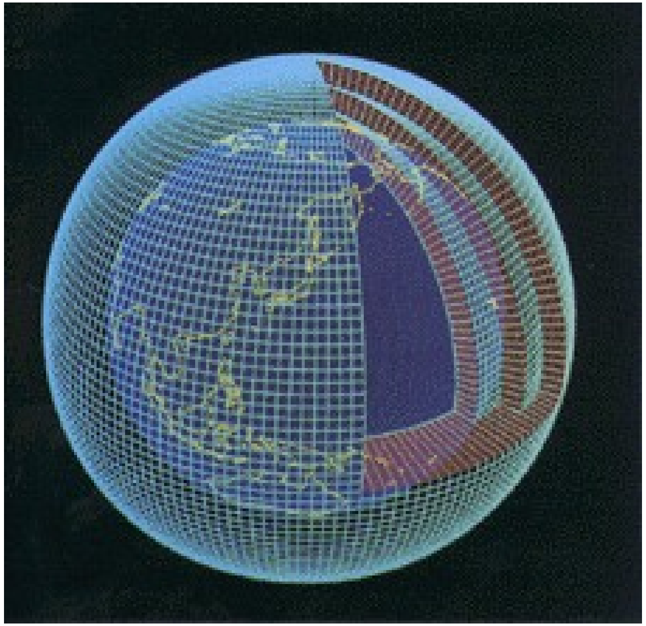
Continuity

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0.$$

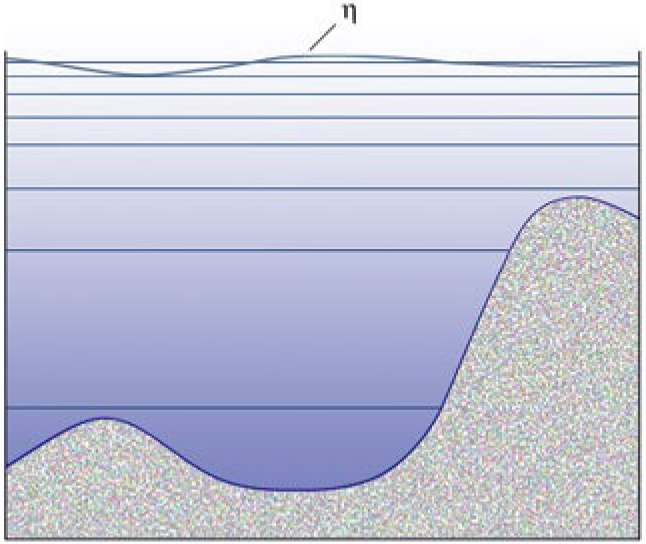
Similar transport equations for other tracers: passive or actives

Horizontal and vertical descretisation of the primitive equations



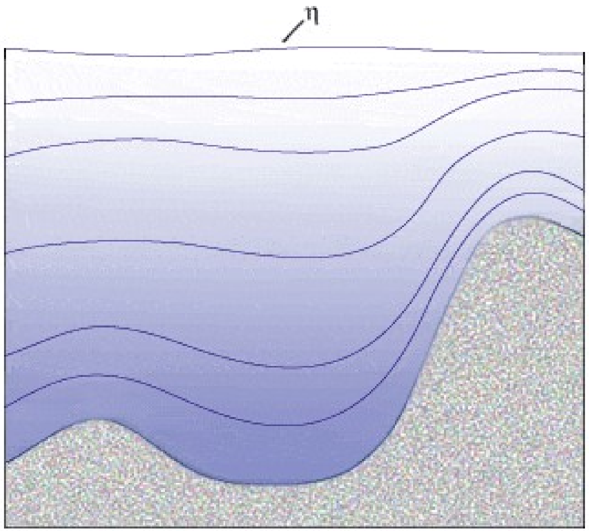


z coordinate system



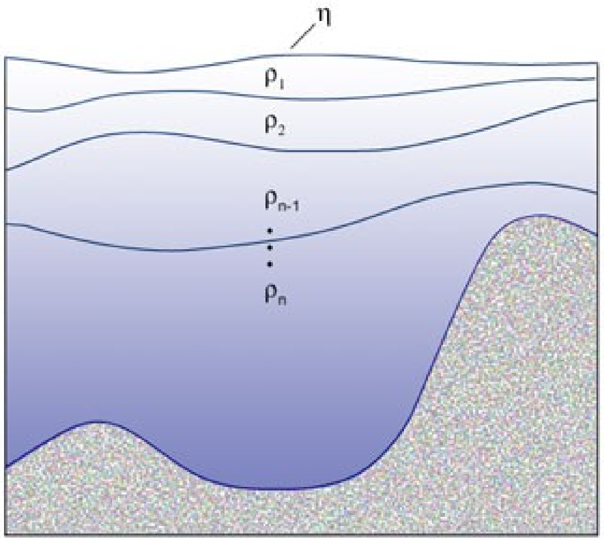
- The vertical coordinate is depth, or "z".
- There are a fixed number of depth levels in the ocean
- This system is the simplest
- z-coordinate models excel in areas that are well-mixed
- z-coordinates have a disadvantage is in regions of sloping topography

Sigma coordinate system



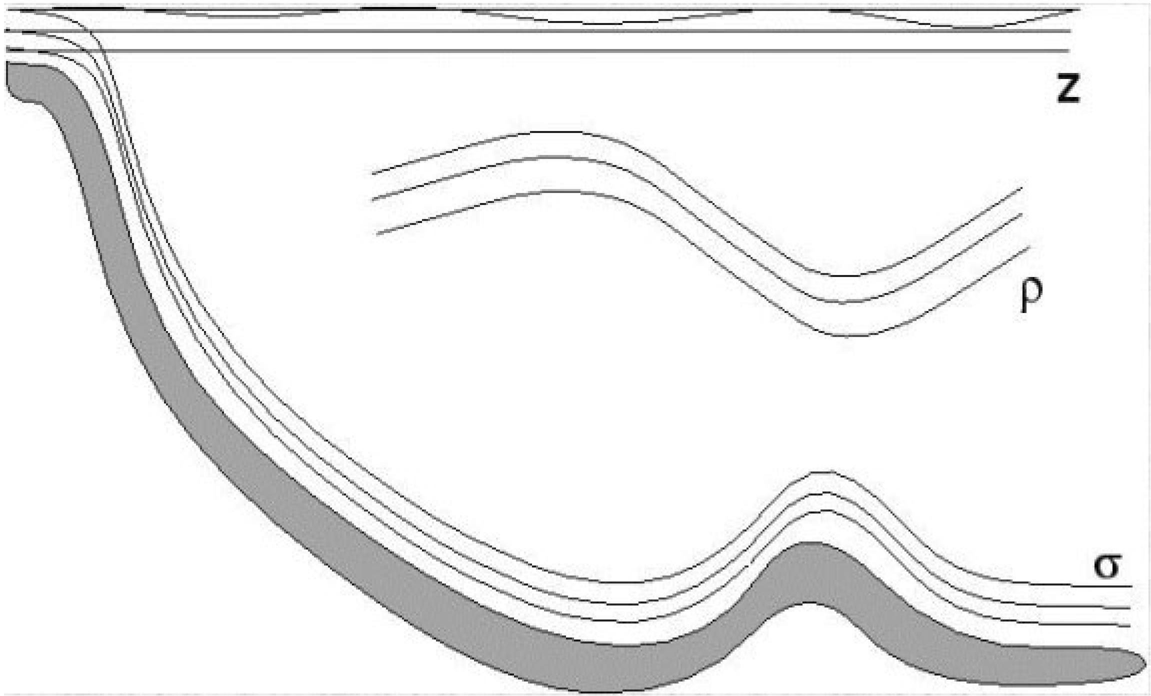
- The vertical coordinate follows the bathymetry
- The layer thicknesses is a function of depth of the ocean
- This type of coordinate is most appropriate for continental shelf and coastal regions
- Pressure-gradient errors can give rise to unrealistic flows where there are sharp gradient in bathymetry

isopycnal coordinate system



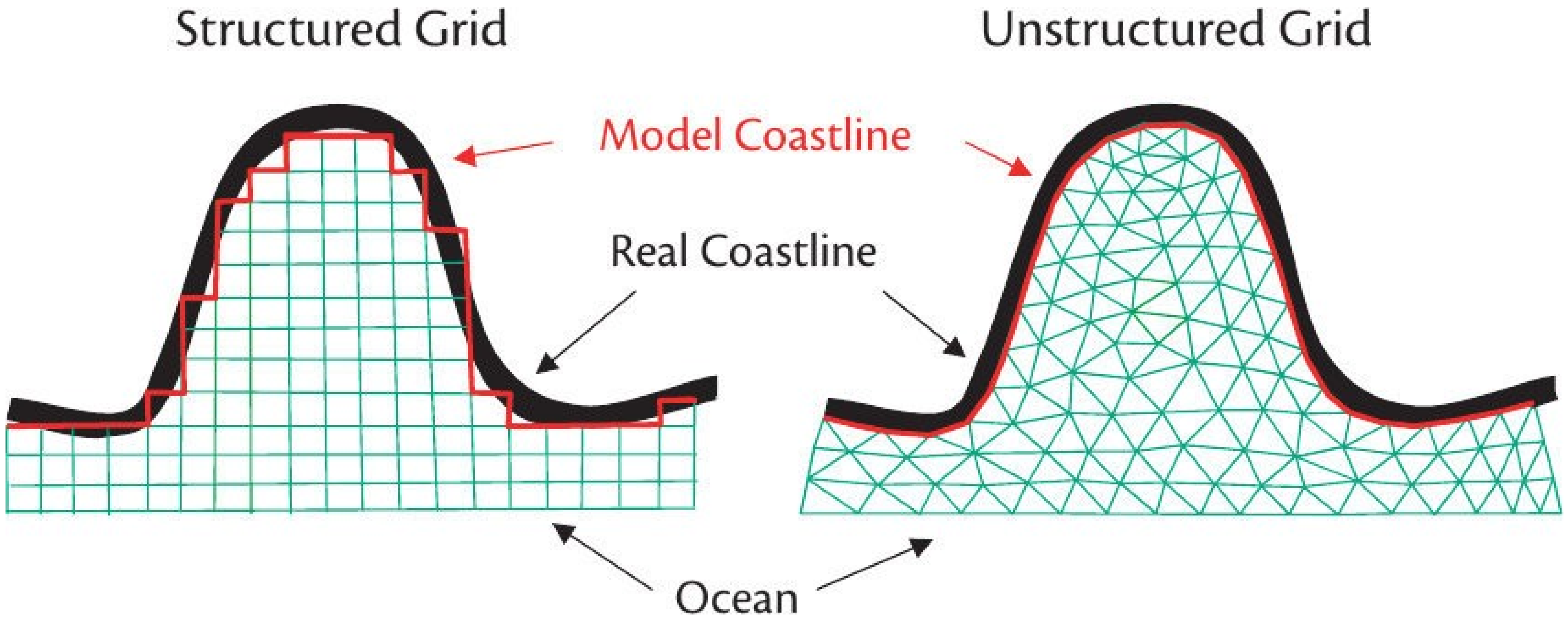
- Use the potential density referenced to a given pressure as the vertical coordinate
- This choice of coordinate works well for modeling tracer transport, which tends to be along surfaces of constant density.
- Because cross-isopycnal mixing is not allowed in this type of model, this type of model has limited applicability in coastal regions and in the surface and bottom boundary layers.

Currently, there are three main vertical coordinates in use, none of which provides universal utility. Hence, many developers have been motivated to pursue research into hybrid approaches.



- Courtesy of Eric Chassignet (U. Miami)

HYCOM and MOM6 make use of hybrid co-ordinate system.



Structured
(MOM, POM, HYCOM,ROMS)

Unstructured
(FVCOM, ADCIRC)

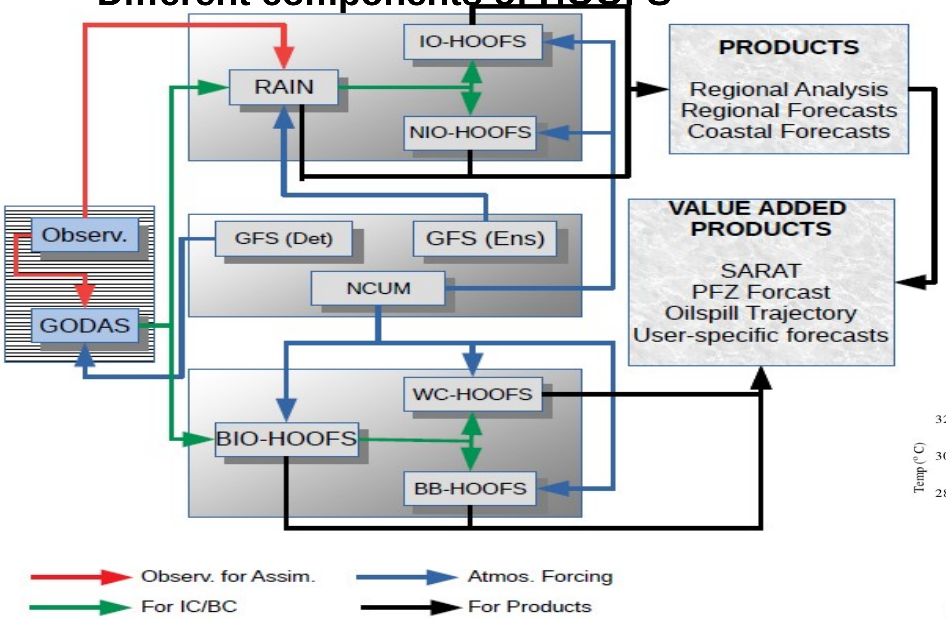
Arakawa Grids (A,B,C,D)

Steps involved in Configuring an Ocean General Circulation Model.

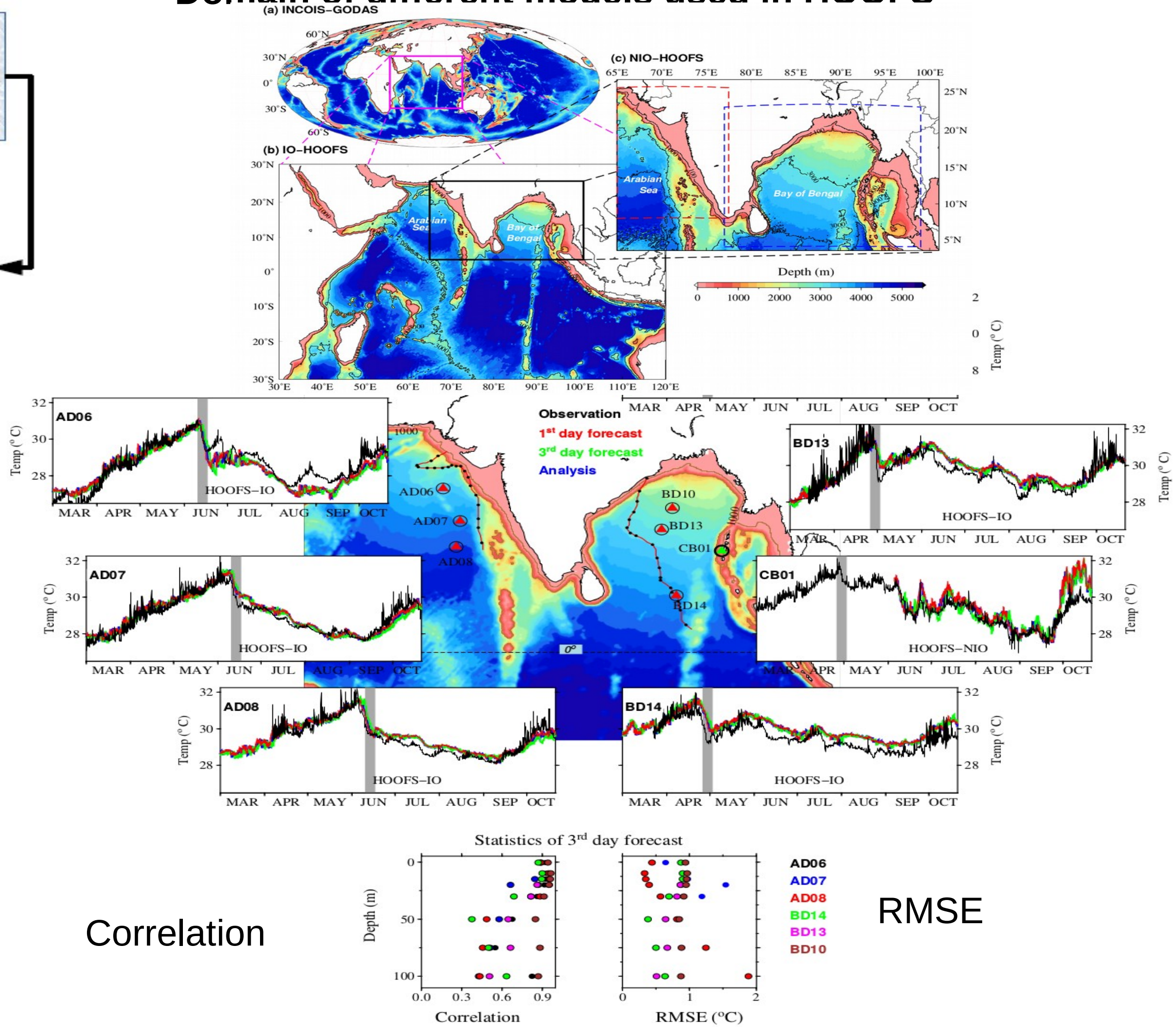
- Define the problem:- For what purpose the model is being configured?
    - Identify the key processes (physical/biogecohemical)
    - Identify the most suitable domain for the model to setup
    - Identify the most suitable co-ordinate system for the purpose
    - Identify the most suitable model
    - Identify the most suitable resolution of the model setup
    - Identify the most suitable forcing parameters
    - Identify the most suitable parameterisation schemes
    - Identify the most suitable initial and boundary conditions
- Download and model porting.
- Prepare the model grid:- Interpolation of the bathymetry into model resolution, open/close the chanels/river mouths/esturies etc.
- Choose the vertical levels/resolution and set the necessary parameters in the model configuration options for this based on the co-ordinate system chosen
- Prepare initial conditions: interpolate to model grid
- Prepare Forcing files
- Prepare boundary conditions
- Specify emperical constants/coefficients
- Choose appropriate time stepping
- Choose appropriate frequency of output
- Integrate model

High-resolution Operational Ocean Forecast and reanalysis System for the Indian Ocean

Different components of HOOFS



Domain of different models used in HOOFS



Correlation

RMSE

- **Developments in**
  - **Physical model**
  - **Biological model**
  - **Data Assimilation**

- **Results from recent research works**

Francis et al., BAMS (2020, under revision)

What we learned today?

1. General overview of climate models
2. Governing Equations
3. Features of Ocean General Circulation
4. Ocean Mixing Processes
5. Representation of mixing processes in the ocean models
6. Horizontal and vertical co-ordinate systems
7. Basic steps involved in configuring an ocean circulation model
8. Brief introduction on the HOOFS

Thank You

For further discussion, pl feel free to contact me at francis@incois.gov.in