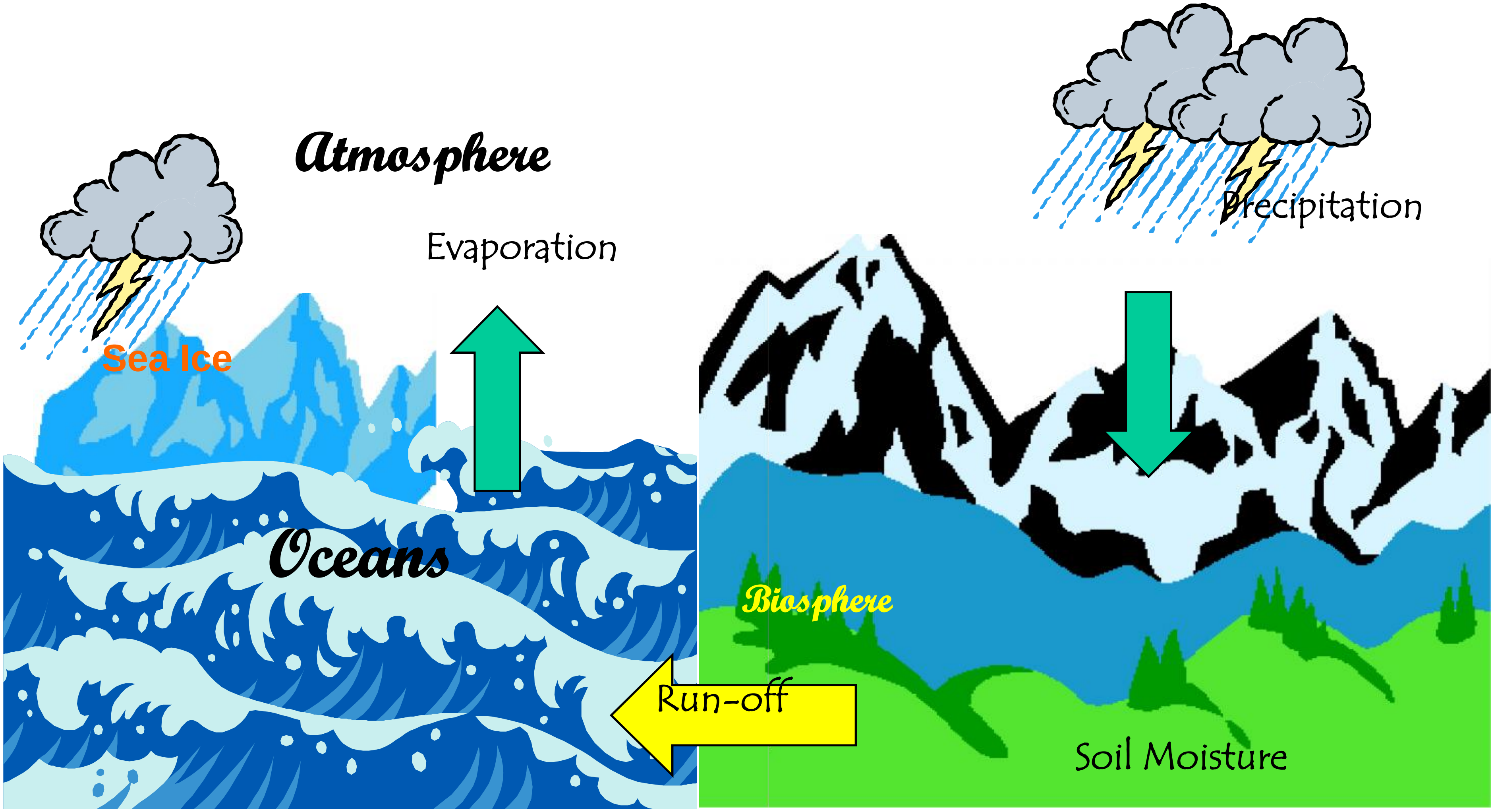
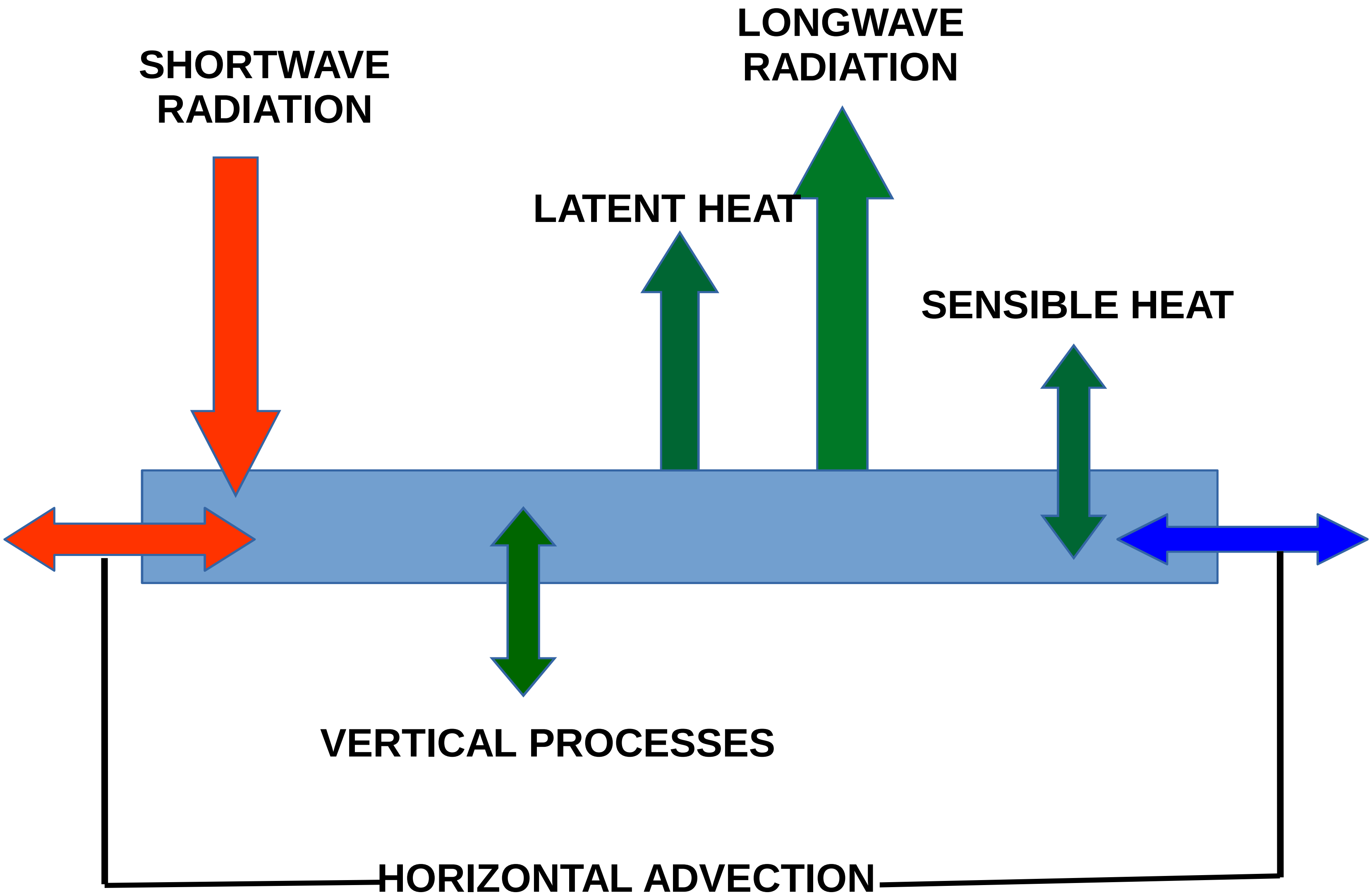
Ocean General Circulation Modeling

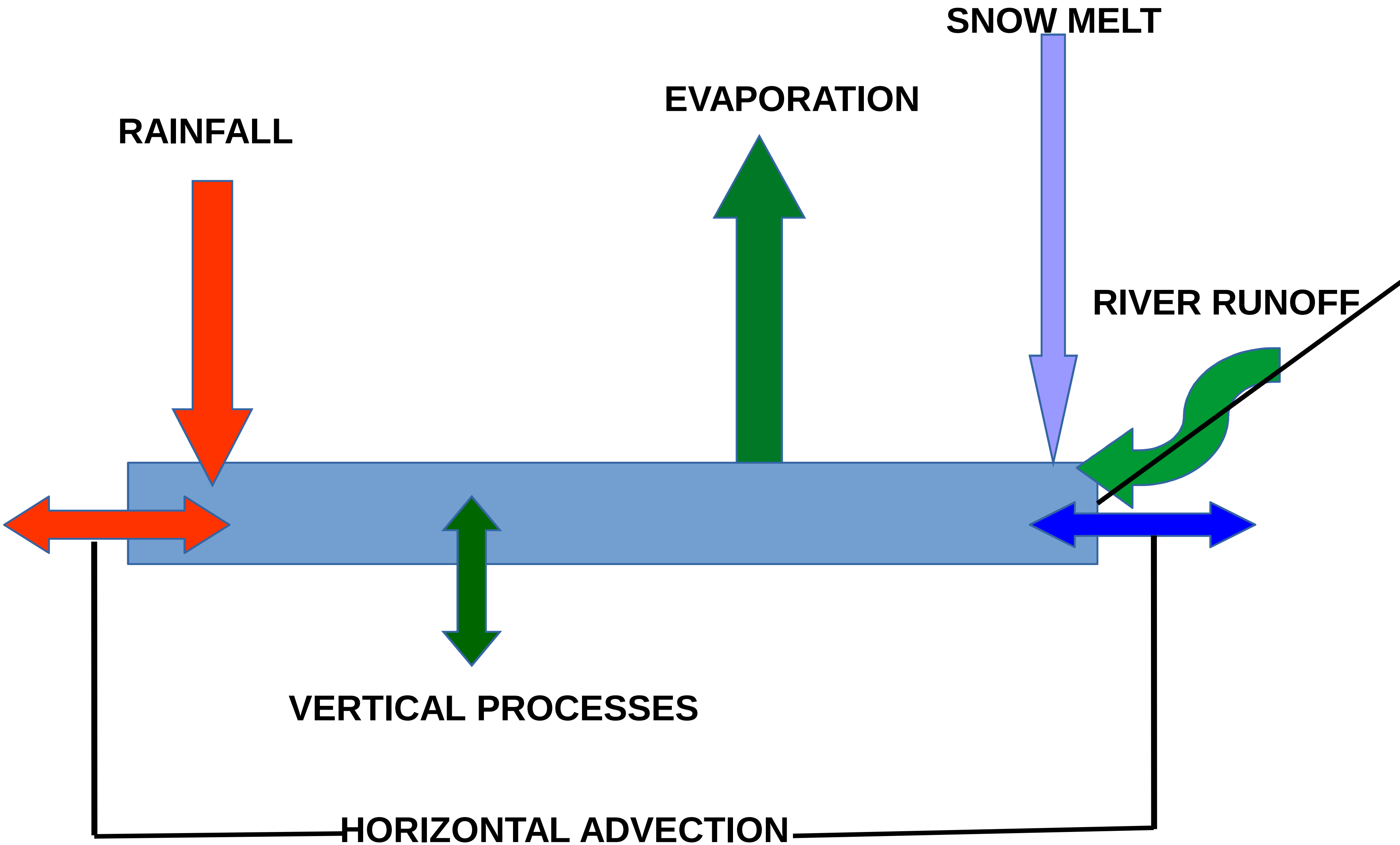
Francis P. A., ESSO-INCOIS



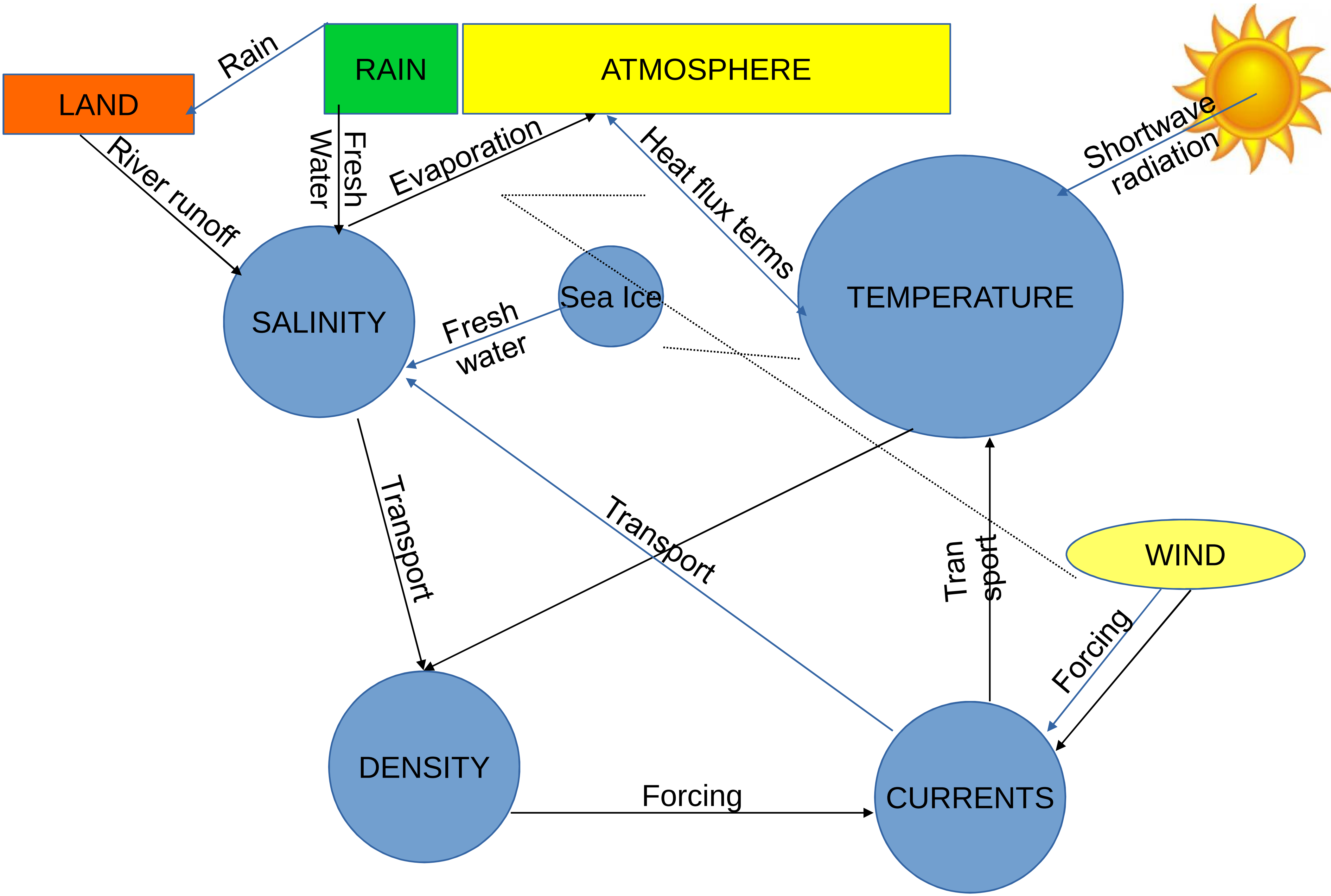
TEMPERATURE



SALINITY



SNOW MELT
EVAPORATION
RAINFALL
RIVER RUNOFF
VERTICAL PROCESSES
HORIZONTAL ADVECTION



LAND
RAIN
ATMOSPHERE
Rain
River runoff
Fresh Water
Evaporation
Heat flux terms
Shortwave radiation
SALINITY
Sea Ice
Fresh water
TEMPERATURE
Transport
Transport
Tran sport
WIND
Forcing
DENSITY
Forcing
CURRENTS

Important Variables

1. Temperature

2. Salinity

3. Velocity

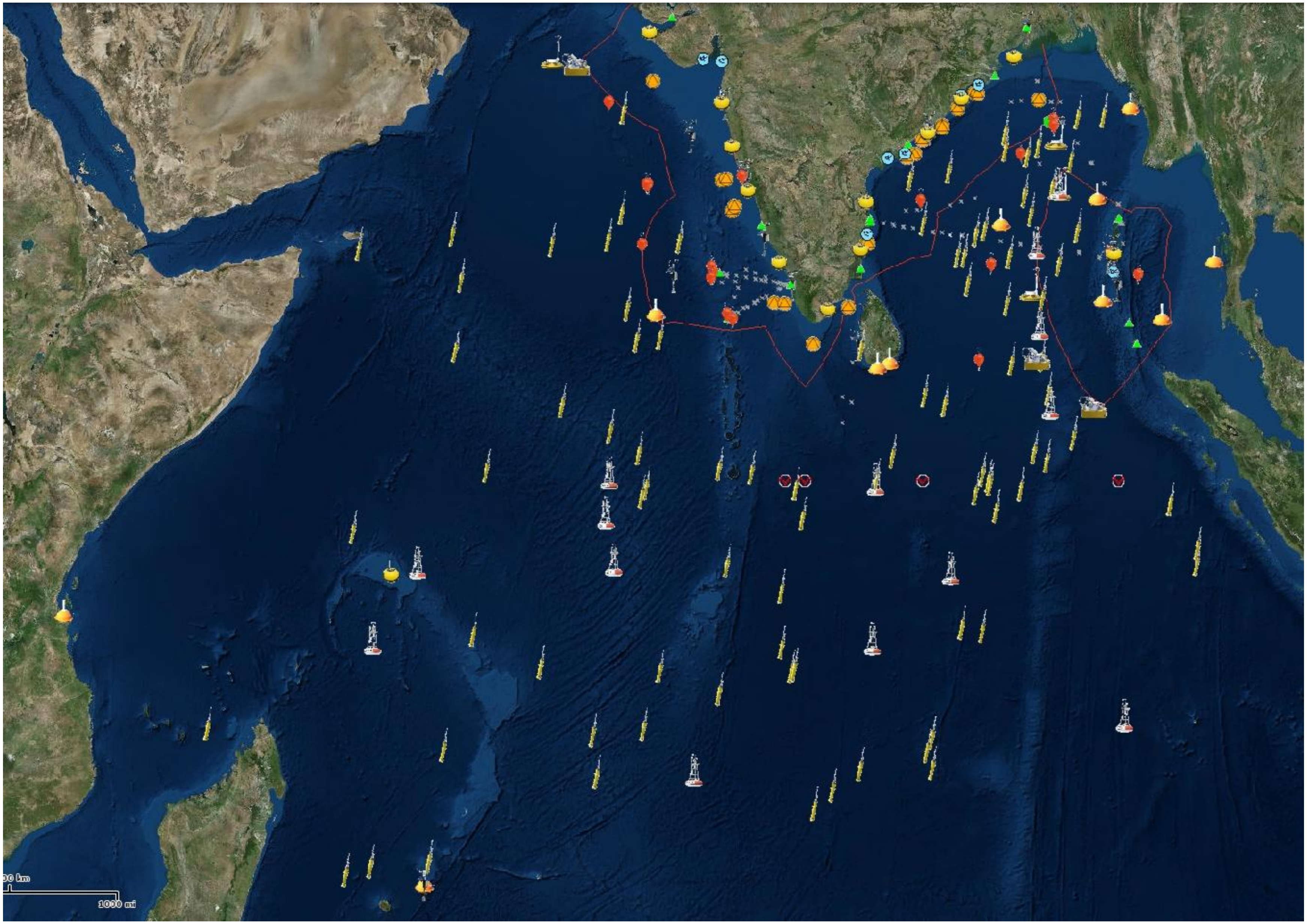
4. Density

5. Pressure

6. Sea surface Hight

7. Dynamic Height

$$D = -\frac{1}{g}\int_{p_0}^{p}\frac{1}{\rho}\,dp$$


1000 mi

General concepts of ocean modeling

What is a numerical ocean model? Why do we need a numerical ocean model?

What is the hierarchy of the ocean models?

How do we design a model?

How do we integrate a model?

What are the equations solved in a numerical circulation model?

What are the possible approximations in an ocean model?

What are the physical processes involved in modeling of upper ocean?

How do we represent the processes in the numerical models?

$$
\begin{aligned}
\partial_t \rho\vec{v} + \nabla\cdot\rho\vec{v}\vec{v} + 2\vec{\Omega}\wedge\rho\vec{v} + g\rho\hat{k} + \nabla p &= \nabla\cdot\vec{\tau} && (9.1)\\
\partial_t\rho + \nabla\cdot\rho\vec{v} &= 0 && (9.2)\\
\partial_t\rho S + \nabla\cdot\rho S\vec{v} &= 0 && (9.3)\\
\partial_t\rho\theta + \nabla\cdot\rho\theta\vec{v} &= \frac{1}{c_{pS}}\nabla\cdot\mathcal{F}_\theta && (9.4)\\
\rho &= \rho(\theta, S, p) && (9.5)
\end{aligned}
$$

Where ρ is the fluid density, $\vec{v}$ is the velocity, p is the pressure, S is the salinity and θ is the potential temperature which add up to seven dependent variables.

The constants are $\vec{\Omega}$ the rotation vector of the sphere, g the gravitational acceleration and c_p the specific heat capacity at constant pressure. $\vec{\tau}$ is the stress tensor and $\mathcal{F}_\theta$ are non-advective heat fluxes (such as heat exchange across the sea-surface).

1. The Boussinesq approximation

Dynamic perturbations in density, are small compared to the background mean density.

In the Boussinesq approximation, which is appropriate for an almost- incompressible fluid, it assumed that variations of density are small, so that in the intertial terms, and in the continuity equation, we may substitute

$$\rho \longrightarrow \rho_0$$

a constant. However, even weak density variations are important in buoyancy, and so we retain variations in ρ in the buoyancy term in the vertical equation of motion.

Implications:

*** No sound waves travel in horizontal if we assume Boussinesq approximation**

***Reduced gravity**

$$D_t\vec{v} + 2\vec{\Omega} \wedge \vec{v} + \frac{g\rho}{\bar{\rho}}\hat{k} + \frac{1}{\bar{\rho}}\nabla p = \frac{1}{\bar{\rho}}\nabla \cdot \vec{\tau}$$

Hydrostatic Approximation

Vertical component of momentum equation (with Boussinesq approximation) is

$$D_t w + 2\Omega \cos\phi v + \frac{g\rho}{\bar{\rho}} + \frac{1}{\bar{\rho}}\partial_z p = \frac{1}{\bar{\rho}}\nabla \cdot \vec{\tau}^w$$

Up on scale analysis, we may reduce it to

$$\frac{g\rho}{\bar{\rho}} + \frac{1}{\bar{\rho}}\partial_z p = 0$$

or

$$\partial_z p = -g\rho$$

The hydrostatic primitive equations

$$
\begin{aligned}
D_t\vec{v}_h + f\hat{k}\wedge\vec{v}_h + \frac{1}{\bar{\rho}}\nabla_h p &= \frac{1}{\bar{\rho}}\nabla\cdot\vec{\tau}_h \\
p &= \int_z^\eta g\rho\,dz \\
\partial_z w &= -\nabla_h\cdot\vec{v}_h \\
D_t S &= 0 \\
D_t\theta &= \frac{1}{\bar{\rho}c_p}\nabla\cdot\mathcal{F}_\theta \\
\rho &= \rho(\theta, S, z) \\
\partial_t\eta &= -\nabla_h\cdot\int_{-H}^{\eta}\vec{v}_h\,dz + (P-E)
\end{aligned}
$$

Try to work it out!
(Clue: integrating the non-divergent continuity equation in the vertical by applying boundary conditions and include the effect of precipiation & evaporation)

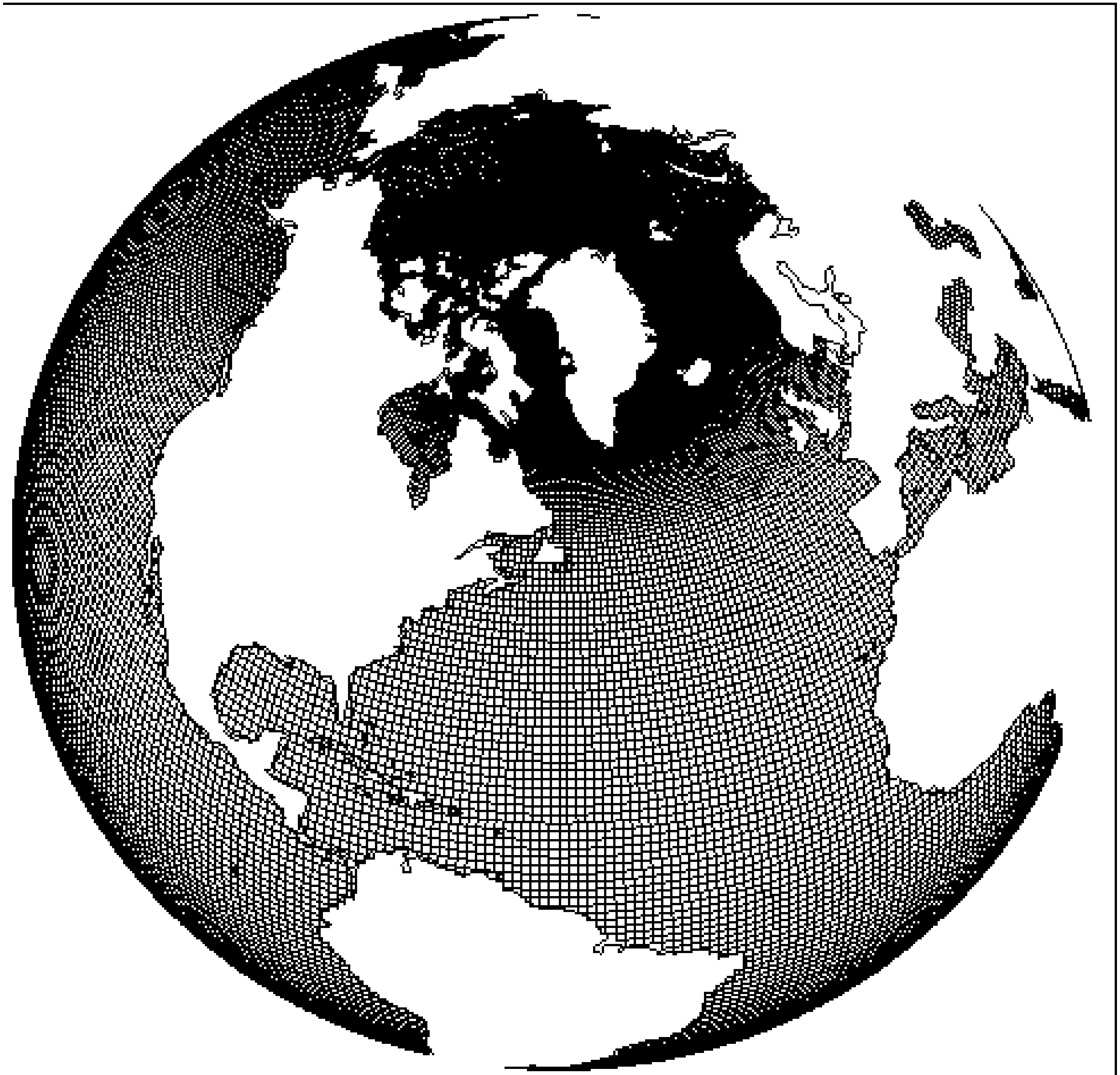
Initial conditions (Data assimilation?)

Boundary conditions

Atmospheric Forcing (heat/momentum/freshwater)

Background emperical constants/coefficients

Integration of the model



Processes in the upper oceans: Mixed Layer Temperature Equation

$$Temperature\,Tendency,$$

$$\partial_t T_{mld} = [Q_{flx}] + [Q_{hadv}] + [Q_{vertical}] + R$$

$$\partial_t T_{mld} = \left[\frac{\left(Q* + Q_S\left(1 - f_{(h)}\right)\right)}{\rho\, C_p\, h}\right]$$

$$- \left[\frac{1}{h}\left(\int_0^h u\,.\,\partial_x T_{mld}\,.\,dz + \int_0^h v\,.\,\partial_y T_{mld}\,.\,dz\right)\right]$$

$$+ \left[\frac{1}{h}\left(\frac{dh}{dt} + w_{(h)}\right)\left(T_{mld} - T_{(h)}\right) + \frac{(K\,\partial_z T)_{(h)}}{h}\right]$$

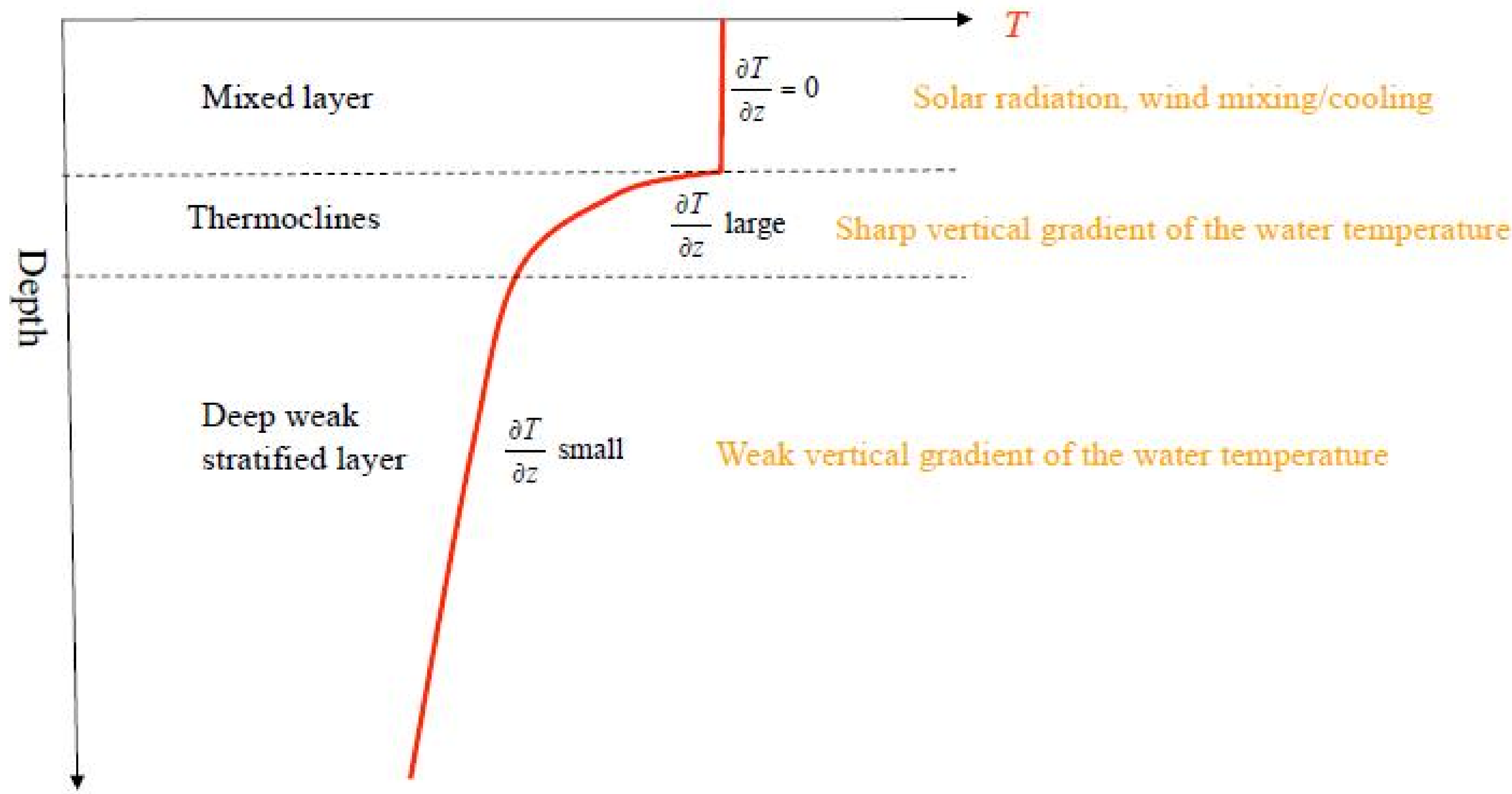
$$+R, \tag{2}$$

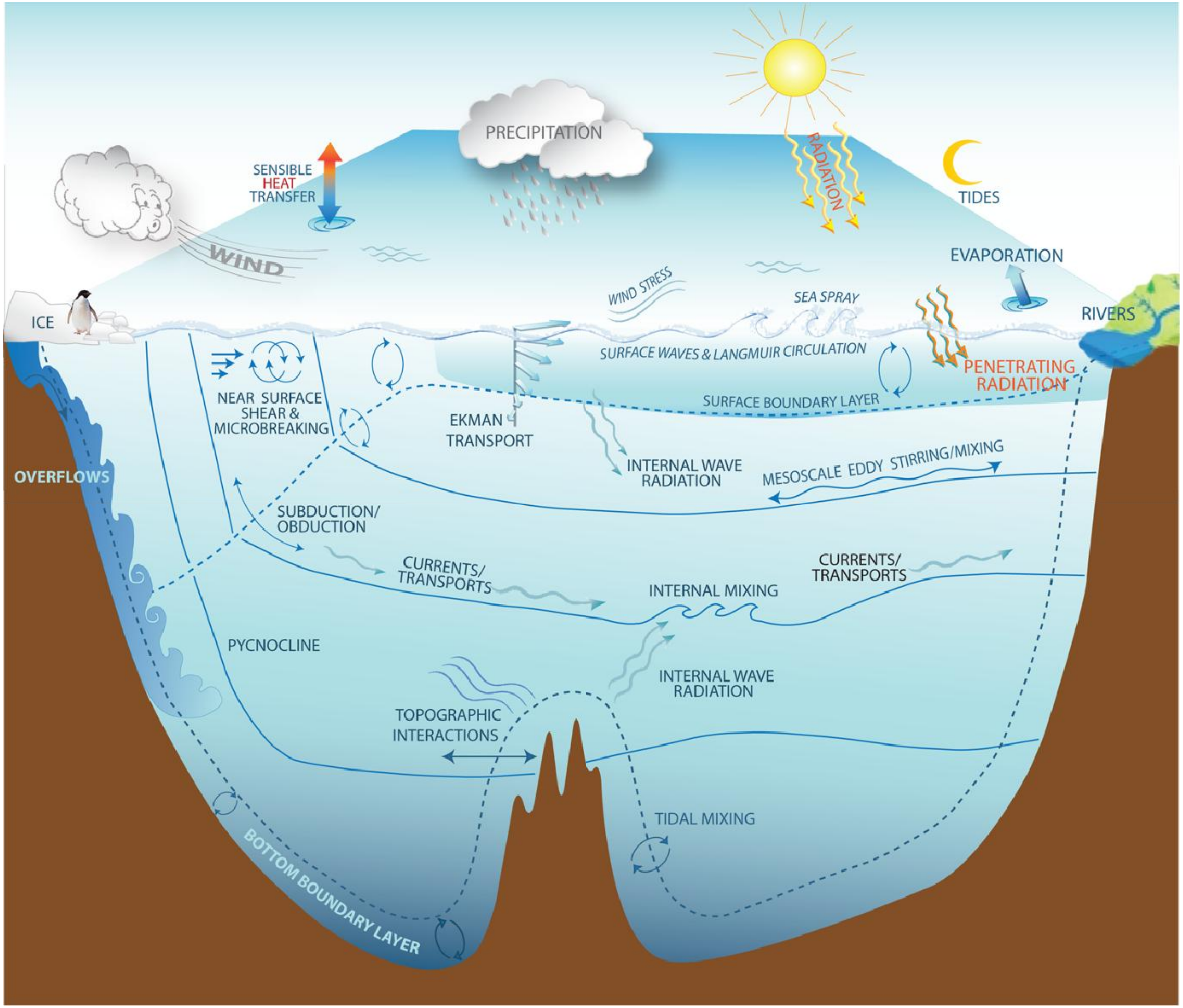
$$Q_{lat} = \rho_a \, L \, C_E \, S \, (q_a - q_s)$$

$$Q_{sen} = \rho_a \, c_{pa} \, C_H \, S \, (T_a - T_s)$$

Q = heat
L = Latent heat of evaporation
C_E= Transfer coefficient of Latent heat
C_H= Transfer coefficient of Sensible heat
c_{pa}= specific heat capacity of air at constant pressure
S= Wind speed
q_a = specific humidity of air
q_s= saturated specific humidity
T_a = air temperature
T_s= SST

The transfer coefficients, C_E and C_H, depend upon the wind speed and stability properties of the atmospheric boundary layer


T
Mixed layer
$\frac{\partial T}{\partial z} = 0$
Solar radiation, wind mixing/cooling
Thermoclines
$\frac{\partial T}{\partial z}$ large
Sharp vertical gradient of the water temperature
Depth
Deep weak stratified layer
$\frac{\partial T}{\partial z}$ small
Weak vertical gradient of the water temperature


PRECIPITATION
SENSIBLE HEAT TRANSFER
RADIATION
TIDES
WIND
EVAPORATION
ICE
WIND STRESS
SEA SPRAY
RIVERS
SURFACE WAVES & LANGMUIR CIRCULATION
PENETRATING RADIATION
NEAR SURFACE SHEAR & MICROBREAKING
EKMAN TRANSPORT
SURFACE BOUNDARY LAYER
INTERNAL WAVE RADIATION
MESOSCALE EDDY STIRRING/MIXING
OVERFLOWS
SUBDUCTION/ OBDUCTION
CURRENTS/ TRANSPORTS
INTERNAL MIXING
CURRENTS/ TRANSPORTS
PYCNOCLINE
INTERNAL WAVE RADIATION
TOPOGRAPHIC INTERACTIONS
TIDAL MIXING
BOTTOM BOUNDARY LAYER

Important mixing mechanisms in the Oceans are

1.Wind and buoyancy-driven mixing in the ocean surface layer,

i). Convection

ii).Wind forcing

a) Breaking waves   b)Langmuir Cells   c) Wind-driven shear d) Temperature ramps

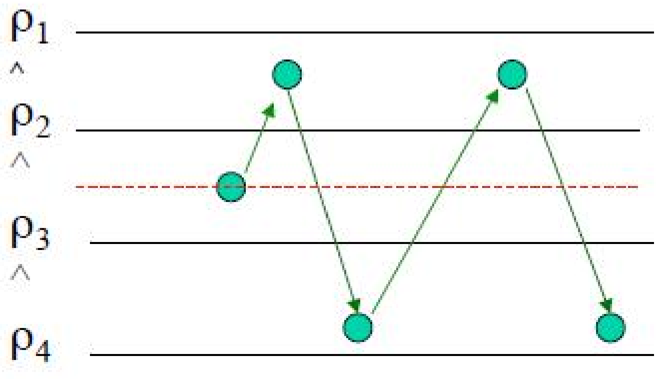
1.Instabilities driven by internal waves in the ocean interior,

2.Double-diffusive interleaving, and

3.Frictionally-driven processes at the bottom boundary/Sea Ice

Note that the horizontal scales of these processes are much smaller than the grid-scales of the climate models.

Turbulance caused by shear instability.

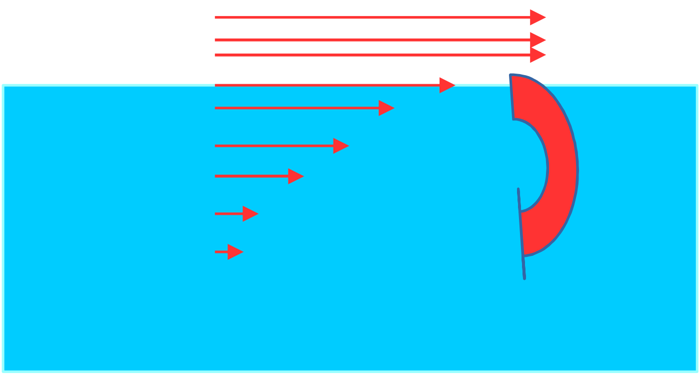


$$F_g = -g(\rho_b - \rho_s)$$

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

→ Brunt-Väisälä frequency

Richardson Number:



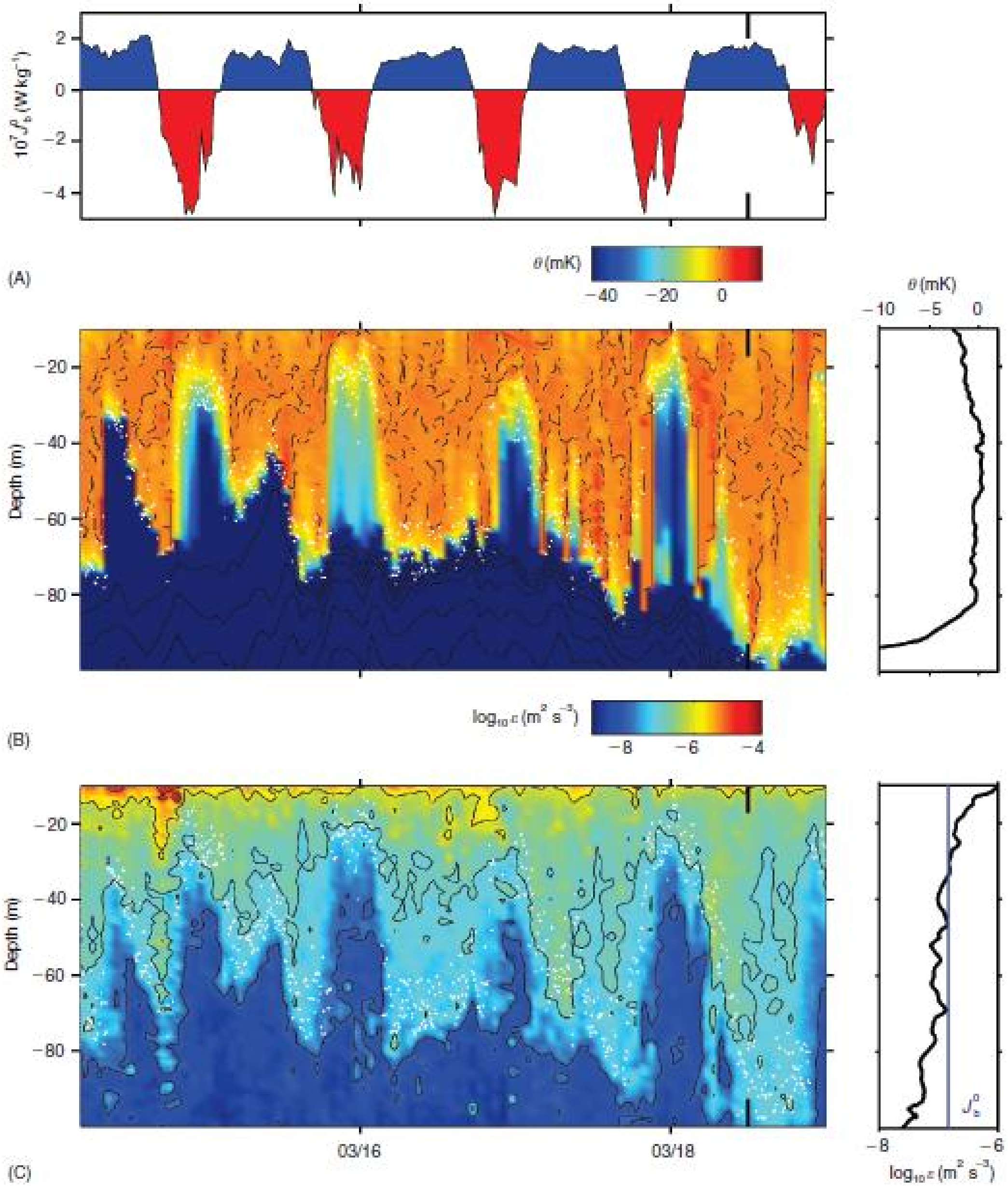
Shear of Velocity

Density stratification

Richardson Number: Ratio of the square of the buoyancy frequency to the square of the vertical shear of the horizontal velocity

$$R_i = N^2 / \left(\frac{\partial u}{\partial z} \right)^2$$

In general, turbulence develops as $R_i < 0.25$, **(critical Richardson number)**
However, critical Richardson number could be bigger for the interior ocean mixing or over the slope.



Courtesy: J. N. Moum and W. D. Smyth, Oregon State University, Corvallis, OR, USA

J^0_b represents the Sux of density (mass per unit volume) across the sea surface due to the combination of heating/cooling and evaporation/precipitation.

(A) The variation in the surface buoyancy flux, J^0_b, which is dominated by surface heating and cooling. The red (blue) areas represent daytime heating (nighttime cooling).
Variations in the intensity of nighttime cooling are primarily due to variations in winds.

(B) Potential temperature referenced to the individual profile mean in order to emphasize vertical rather than horizontal structure (K). To the right is an averaged vertical
profile from the time period indicated by the vertical bars at top and bottom of each of the left-hand panels.

(C) The intensity of turbulence as indicated by the viscous dissipation rate of turbulence kinetic energy,   . To the right is an averaged profile with the mean value of J^0_b indicated by the vertical blue line.

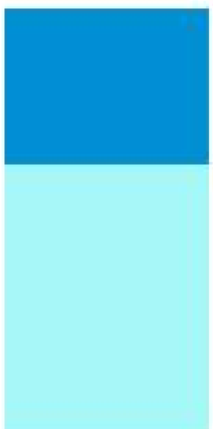
The dots in (B) and (C) represent the depth of the mixed layer as determined from individual profiles

Double Diffusivity

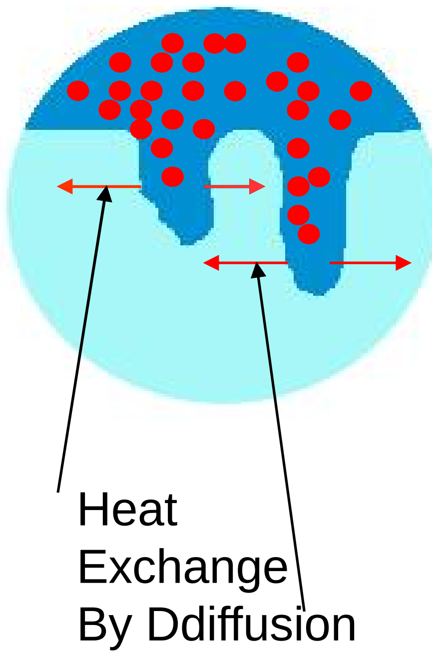
Occurs when the diffsivity of the propreties are not uniform

In the oceans,temperature diffuses approximately 100 times faster than salt!

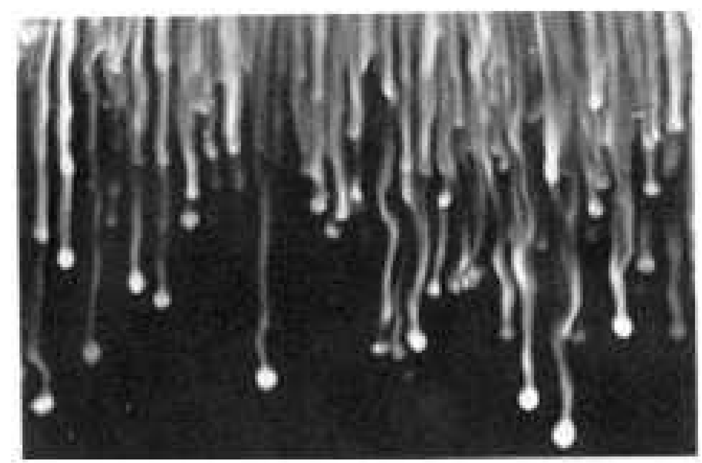
Warm
Salty



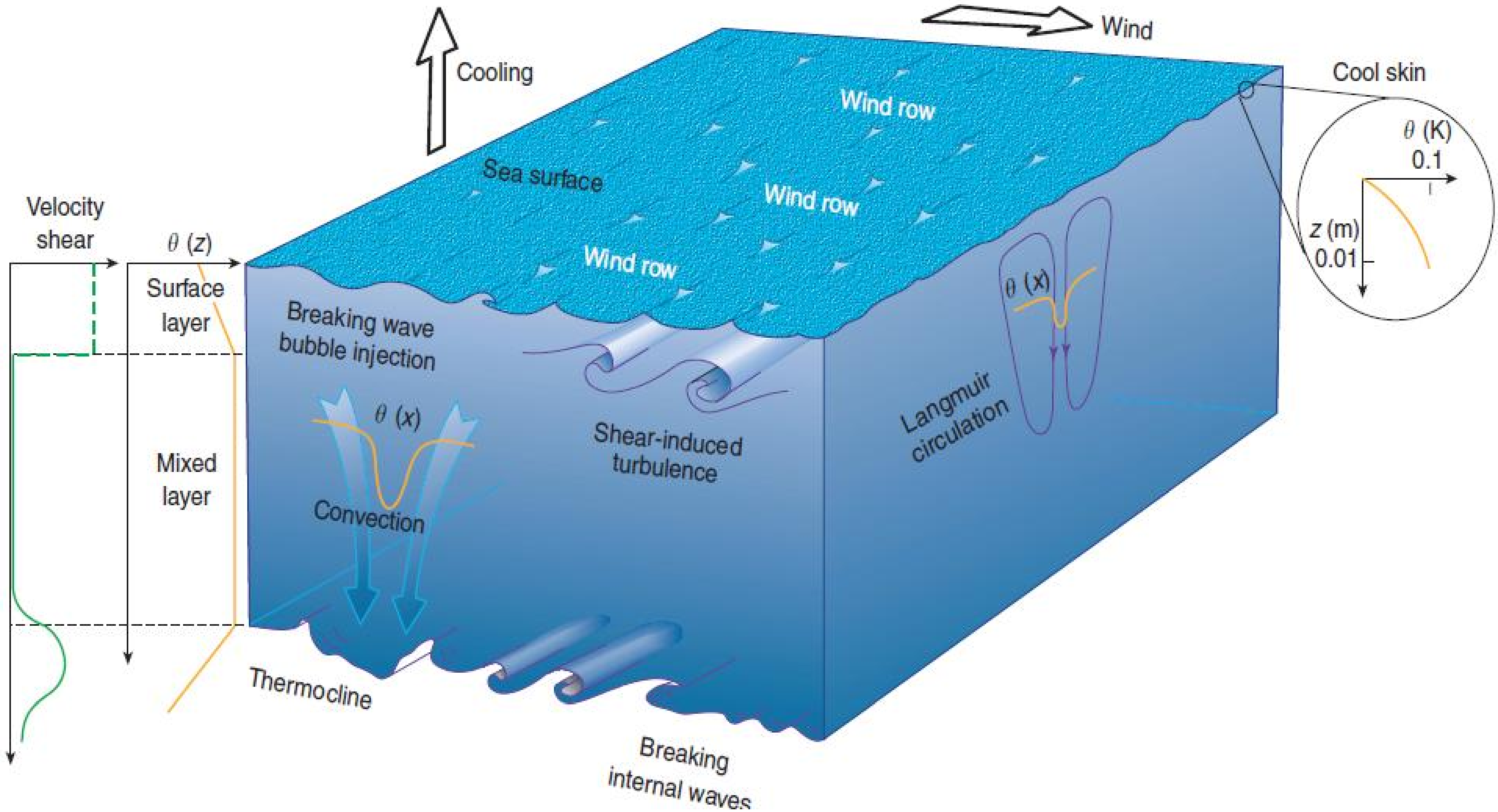
Cold
Fresh



Salt Fingering.

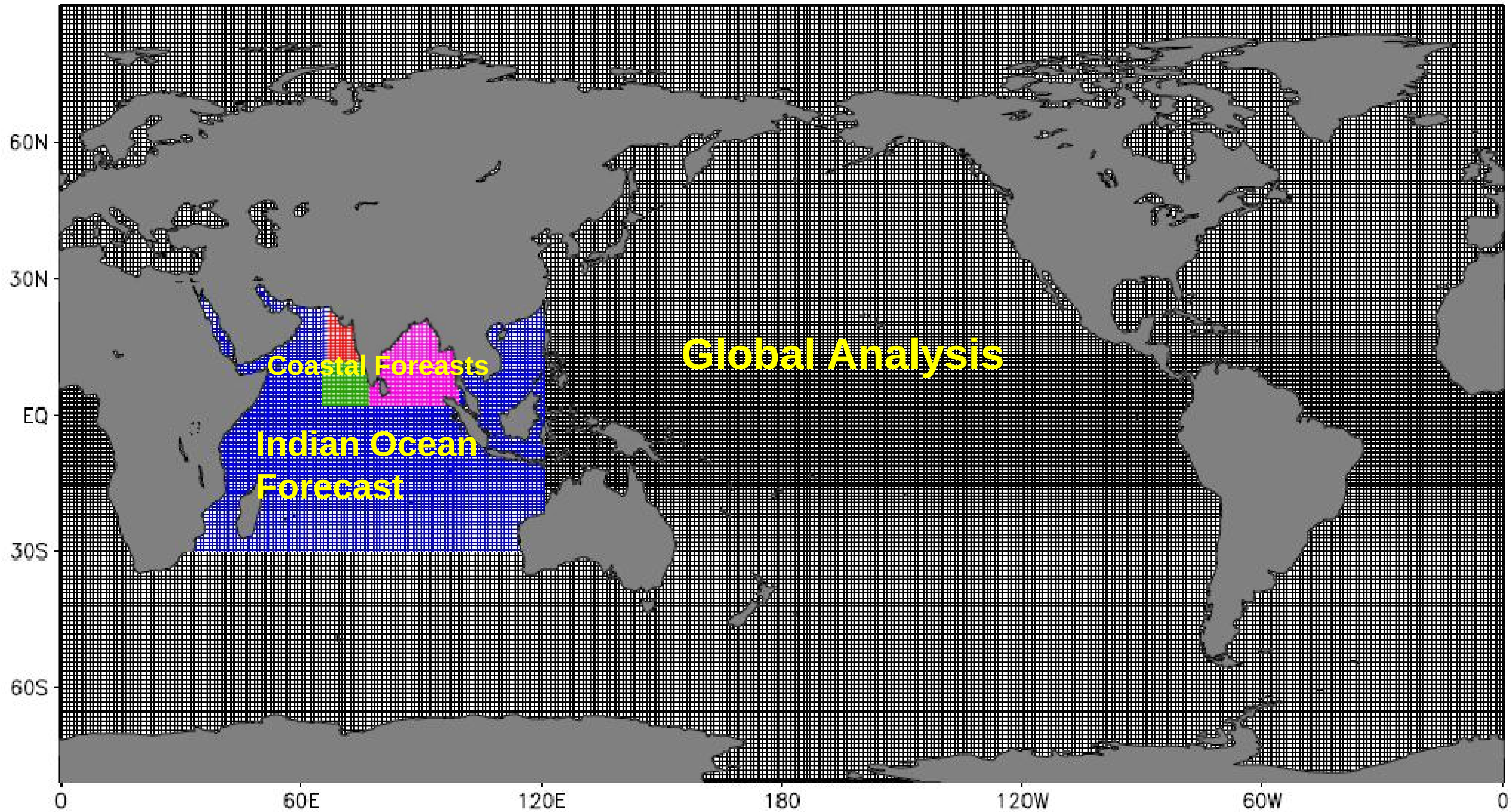


Salt fi
ngers in a laboratory experiment
Courtesy: Timour Radko, WHGFDP



Wind
Cooling
Cool skin
Wind row
θ (K)
0.1
Sea surface
Wind row
Velocity
shear
θ (z)
Wind row
z (m)
0.01
Surface
layer
θ (x)
Breaking wave
bubble injection
θ (x)
Langmuir
circulation
Shear-induced
turbulence
Mixed
layer
Convection
Thermocline
Breaking
internal waves

Architecture of HOOFS- General Circulation Models



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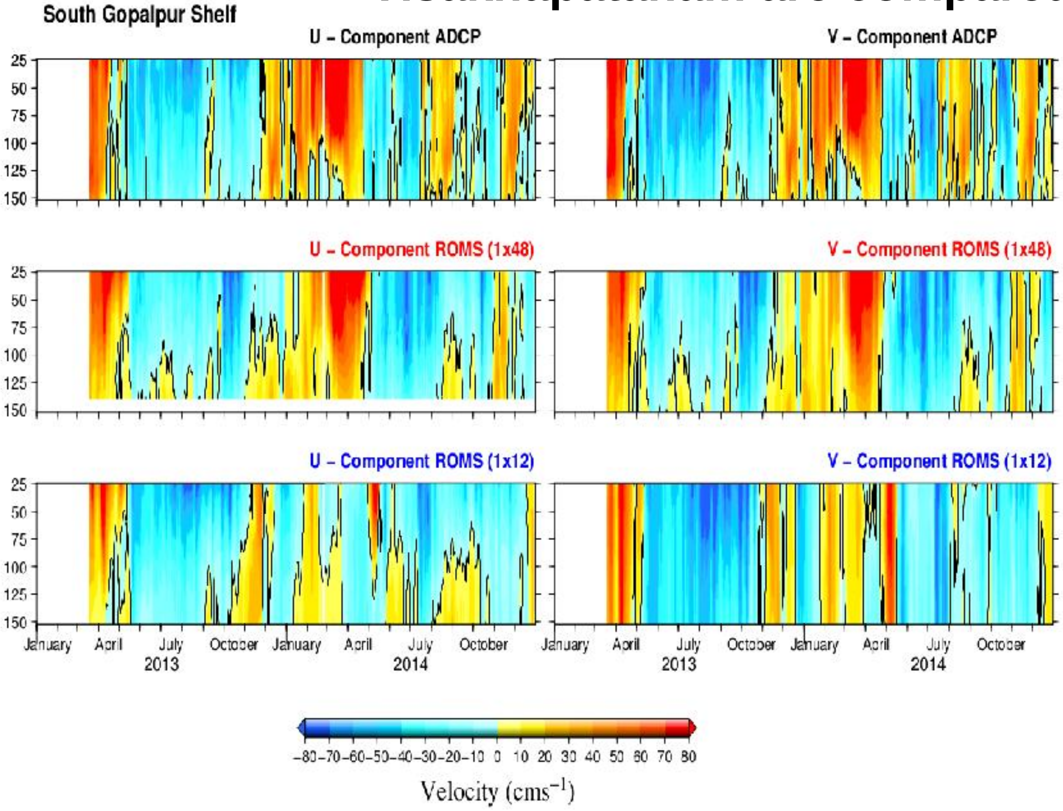
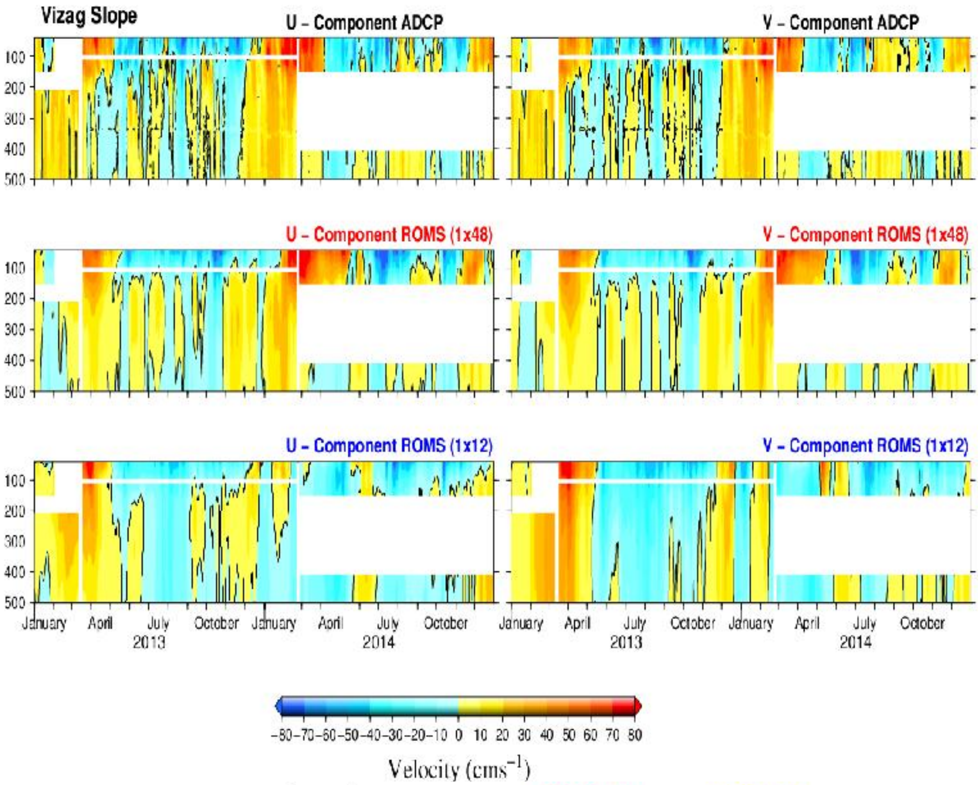
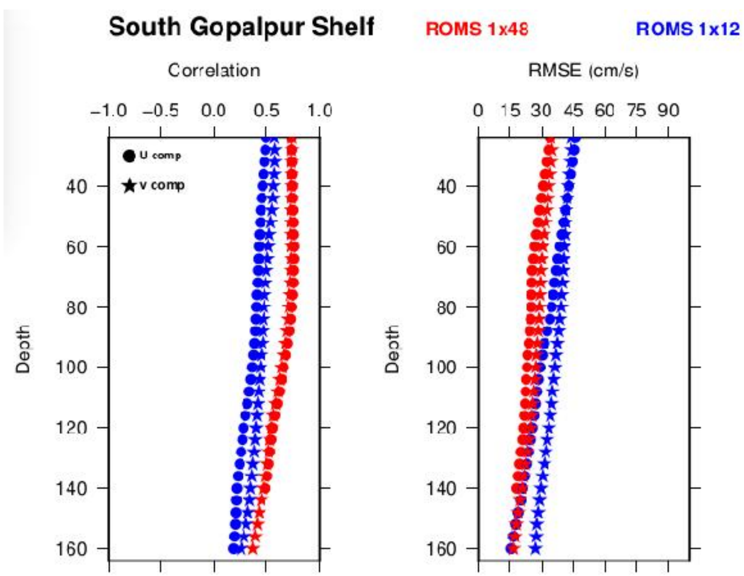
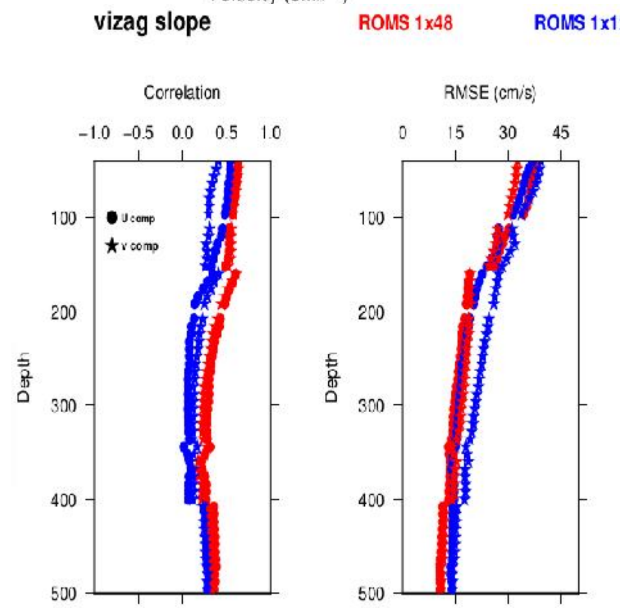
Global : GODAS (variable resolution, MOM4p0d + 3DVAR)
Indian Ocean : ROMS (~9.5 x 9.5 km) & HyCOM with red. KF(~ 6.5 x 6.5 km, experimental)

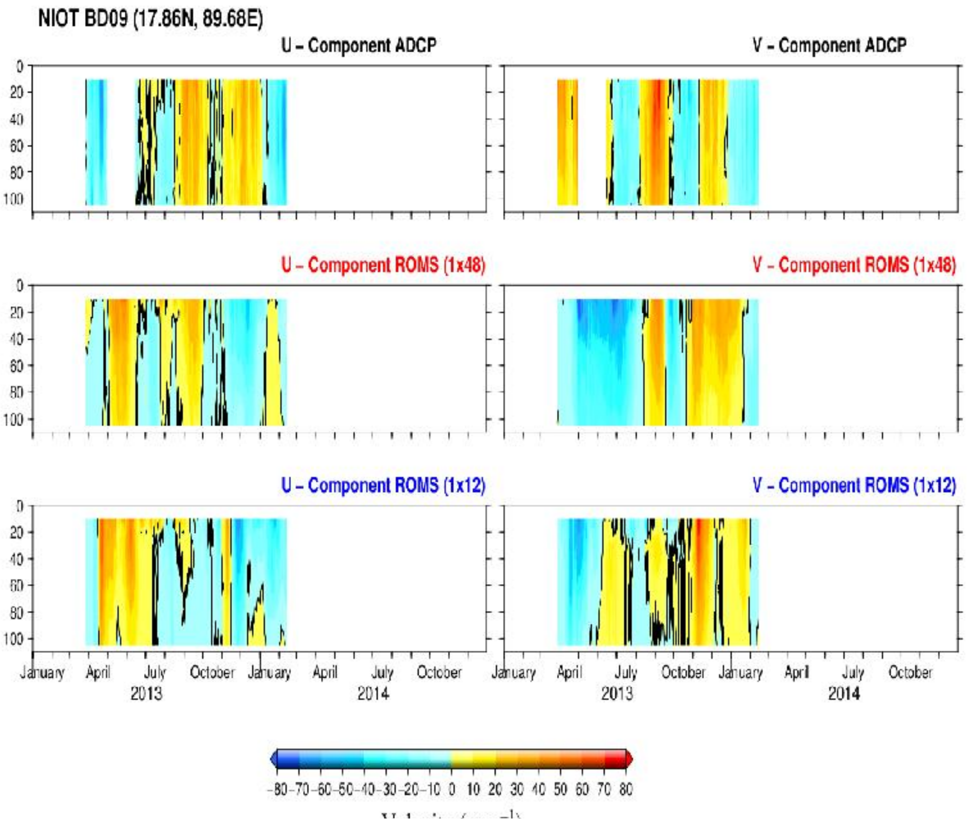
Time-depth section of temperature currents simulated by BB-HOOFS off Gopalpur and Visakhapatanam are compared with ADCP observations



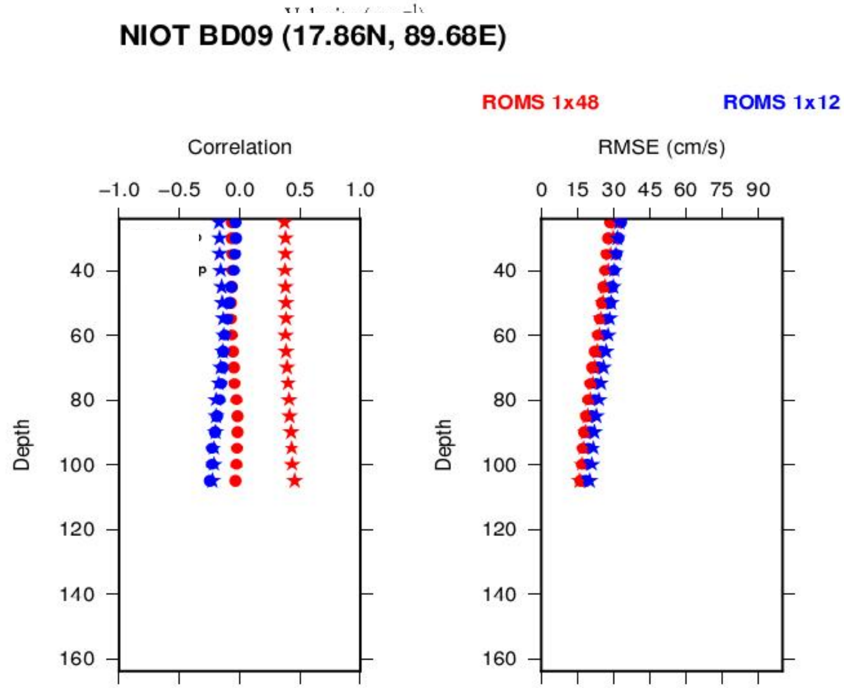




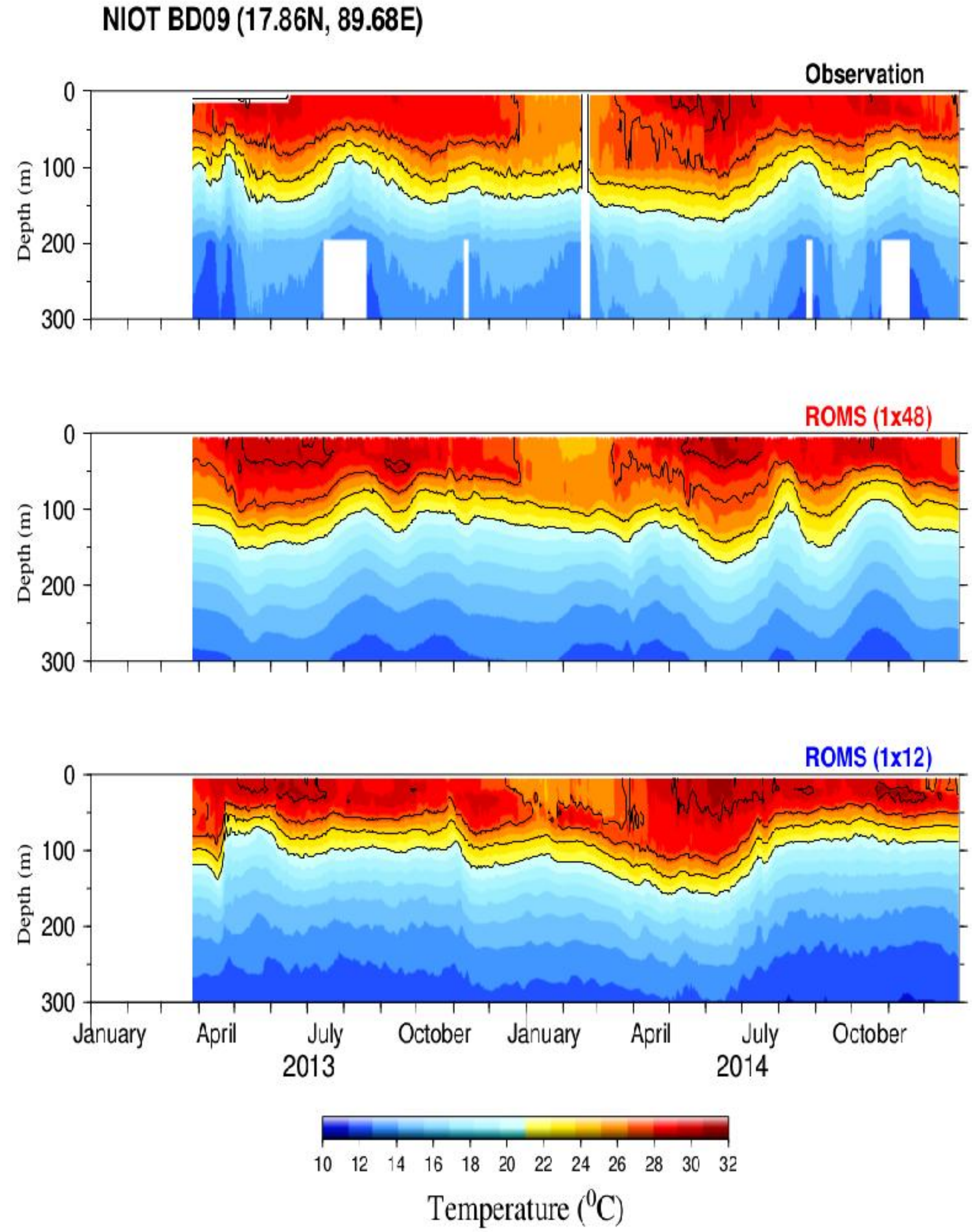




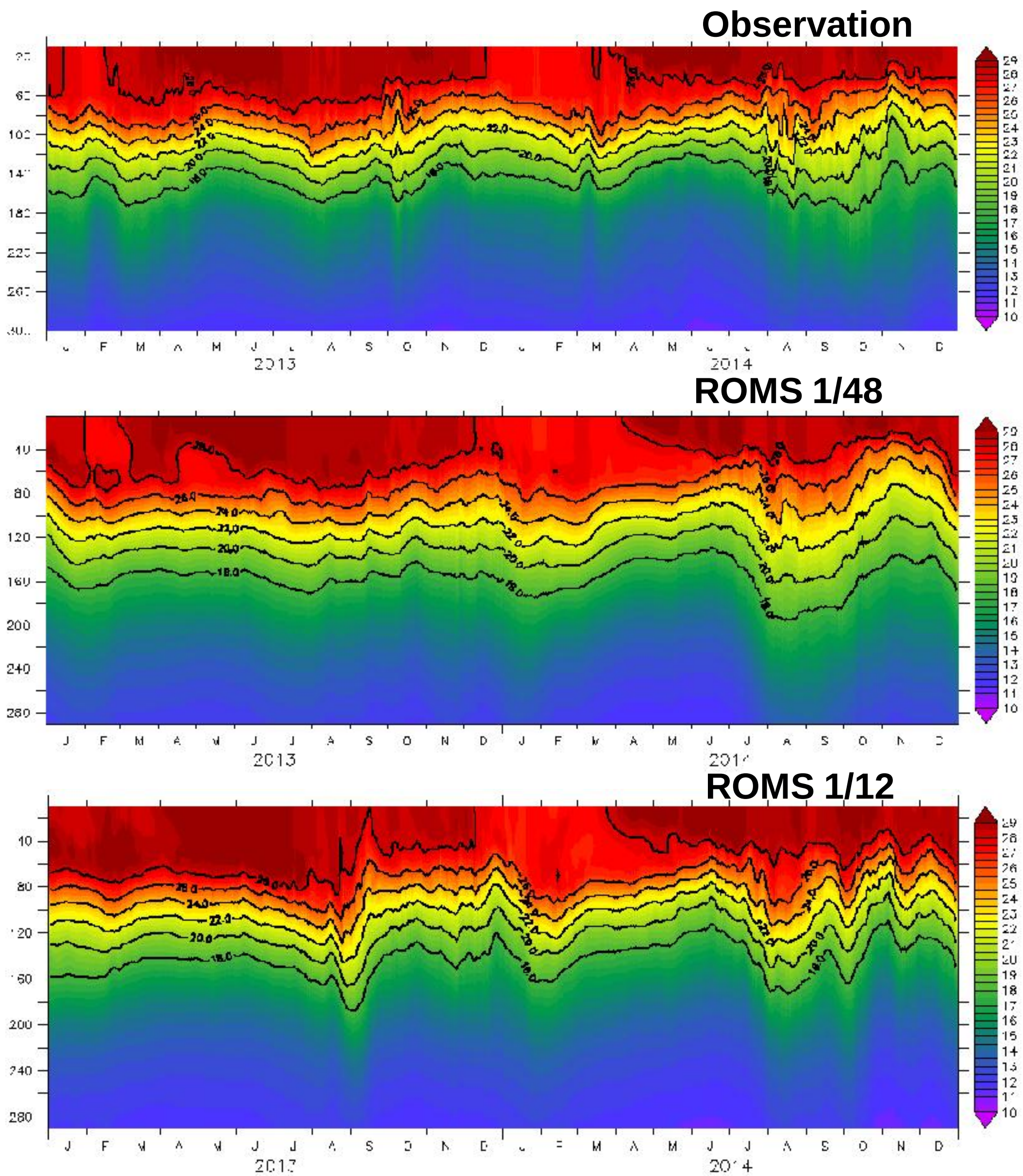
NIOT BD09 (17.86N, 89.68E)
U – Component ADCP
V – Component ADCP
U – Component ROMS (1x48)
V – Component ROMS (1x48)
U – Component ROMS (1x12)
V – Component ROMS (1x12)
January April July October January April July October
2013 2014
-80 -70 -60 -50 -40 -30 -20 -10 0 10 20 30 40 50 60 70 80

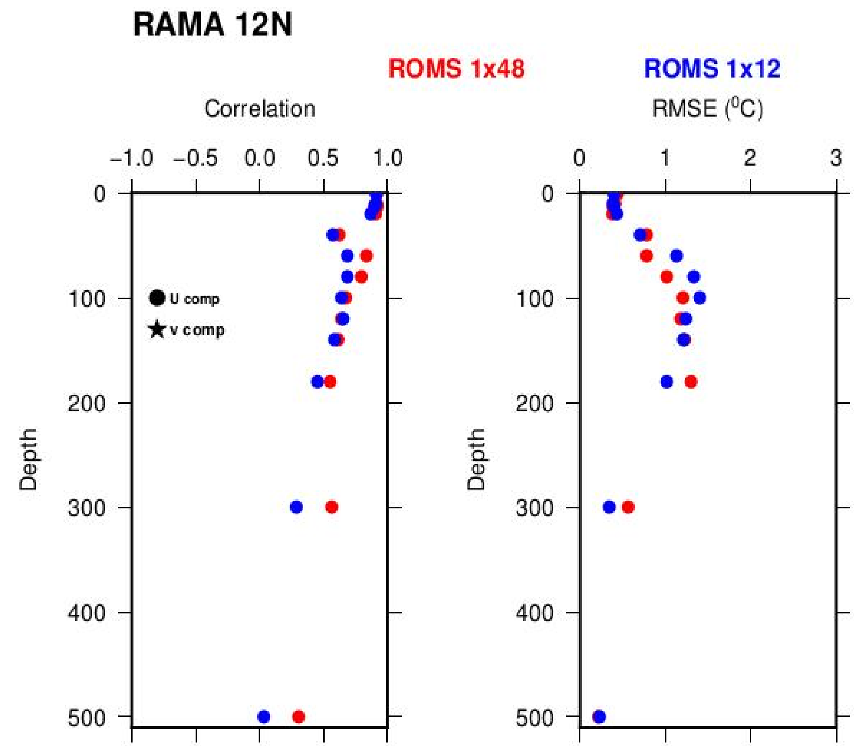


NIOT BD09 (17.86N, 89.68E)
ROMS 1x48
ROMS 1x12
Correlation
RMSE (cm/s)
Depth

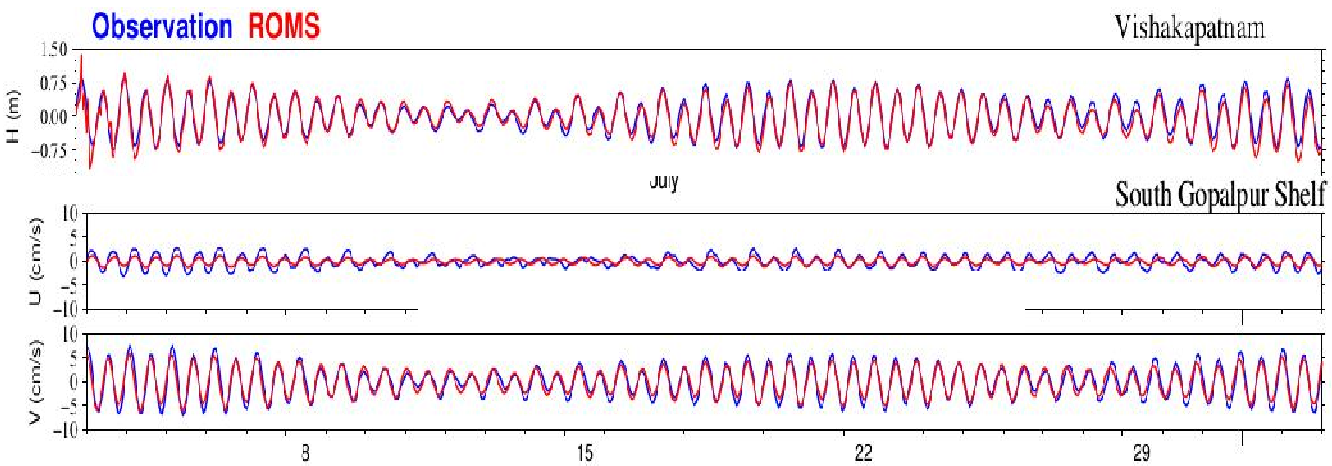


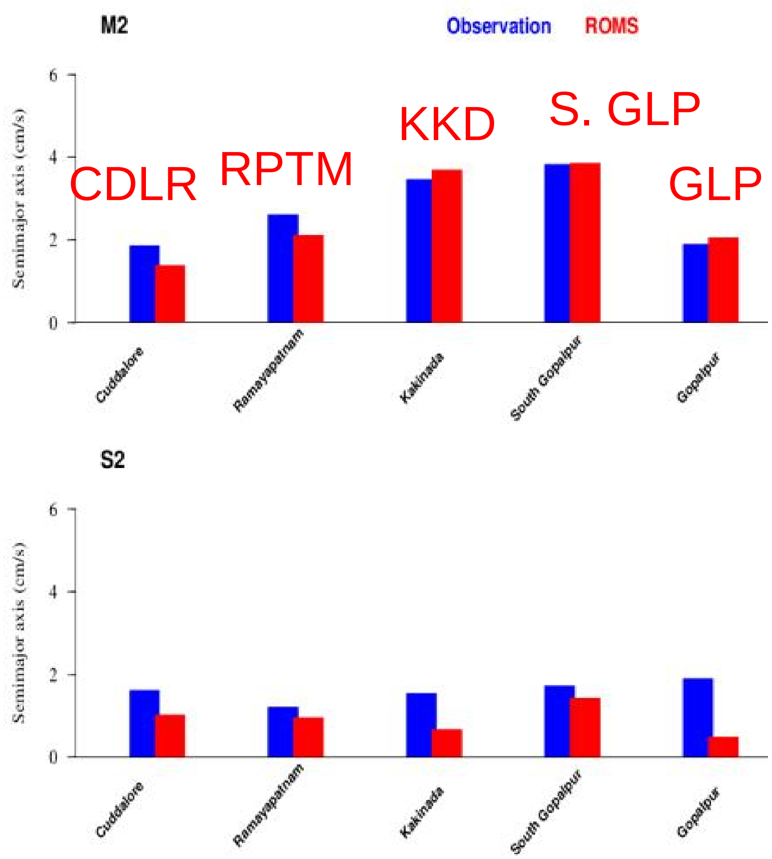
NIOT BD09 (17.86N, 89.68E)
Observation
ROMS (1x48)
ROMS (1x12)
Depth (m)
January April July October January April July October
2013 2014
Temperature (0C)


Observation
ROMS 1/48
ROMS 1/12
2013
2014


RAMA 12N
ROMS 1x48
ROMS 1x12
Correlation
RMSE (⁰C)
Depth
U comp
v comp

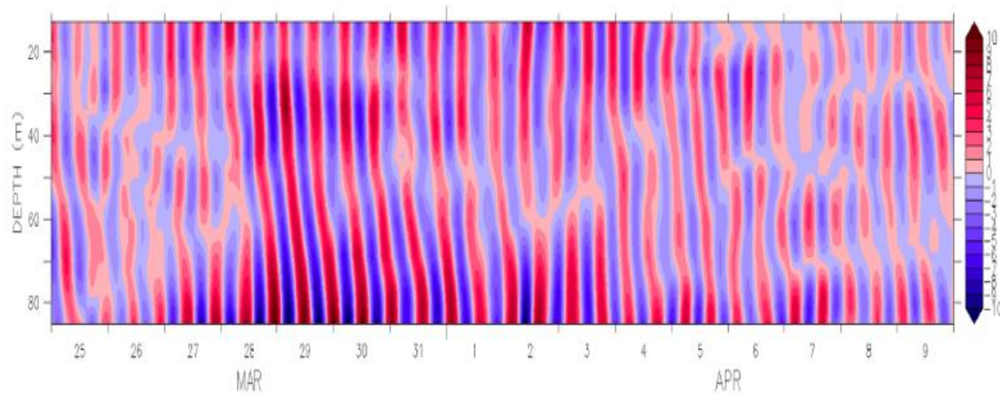
Modeling of tidal currents off the east coast of India using ROMS



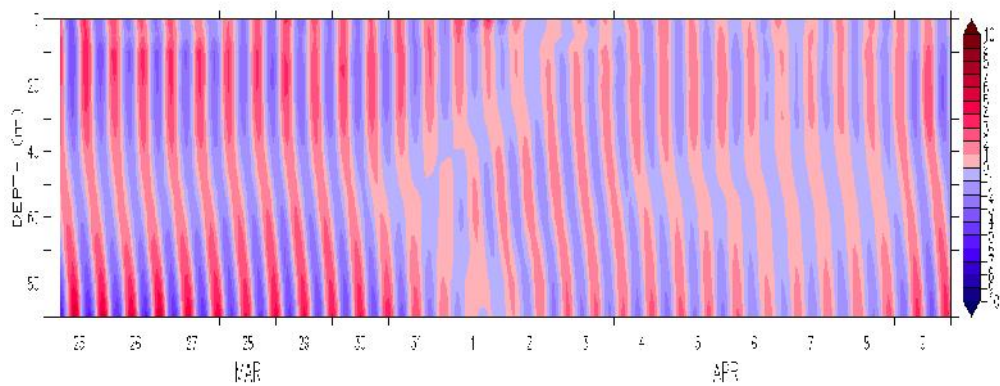


Cross Isobath velocity of semidiurnal internal tide

ADCP

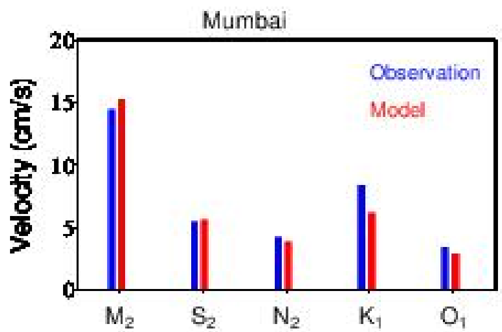


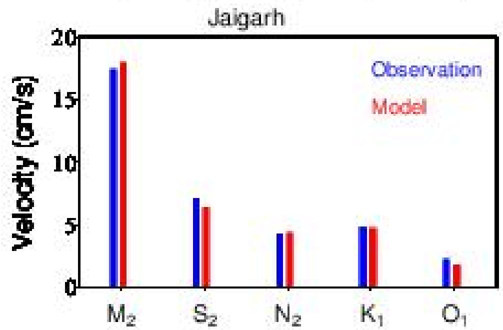
ROMS

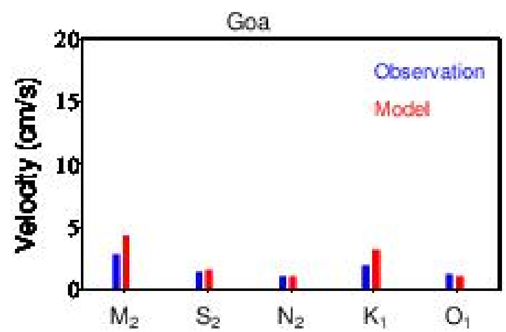


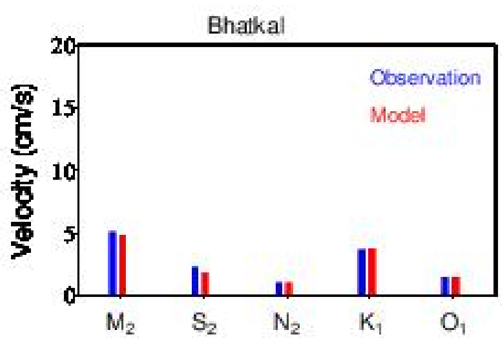
Comparison of barotropic tidal currents on the shelf off the west coast

- ADCP data (Subeesh et al., 2013) for different year is used for this comparison with ROMS simulations.
- Amplitude and phase of tidal constituents do not vary with season so that we can use any year data for comparison.
- Maximum velocity (semimajor axis of ellipse) of 5 constituents is shown

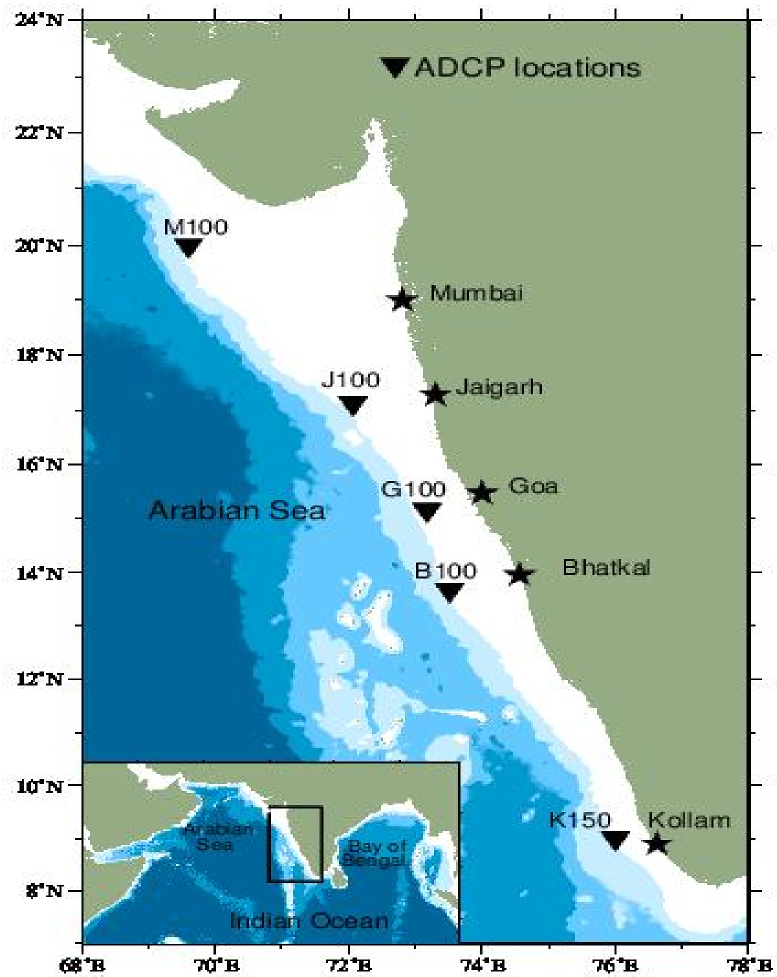


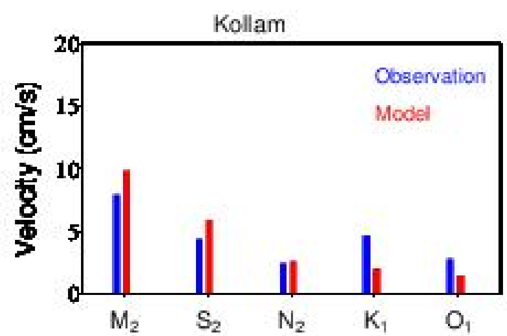


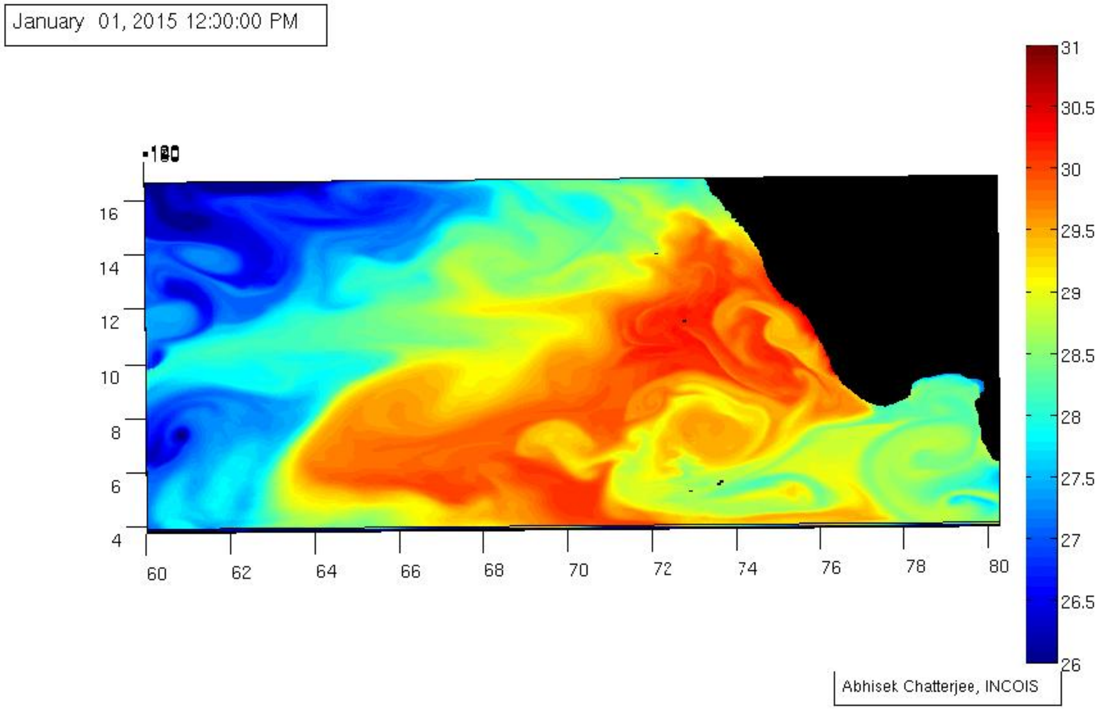




Shelf ADCP locations



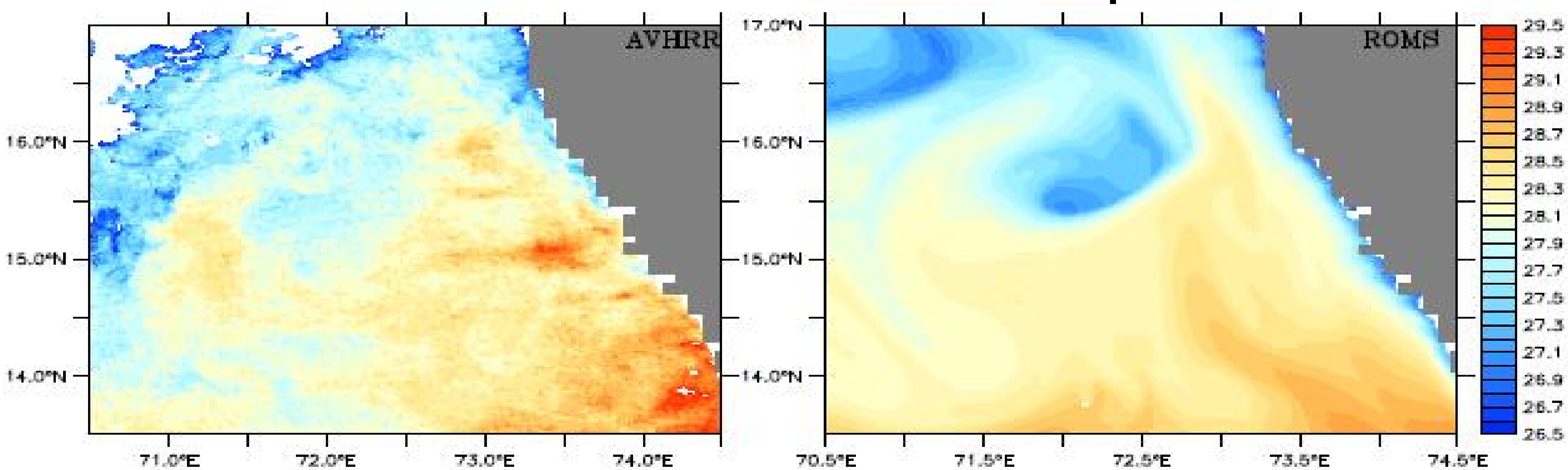




Though the models use state-of the art parameterisation schemes (such as mixing), there is considerable scope for improvements in these schemes.

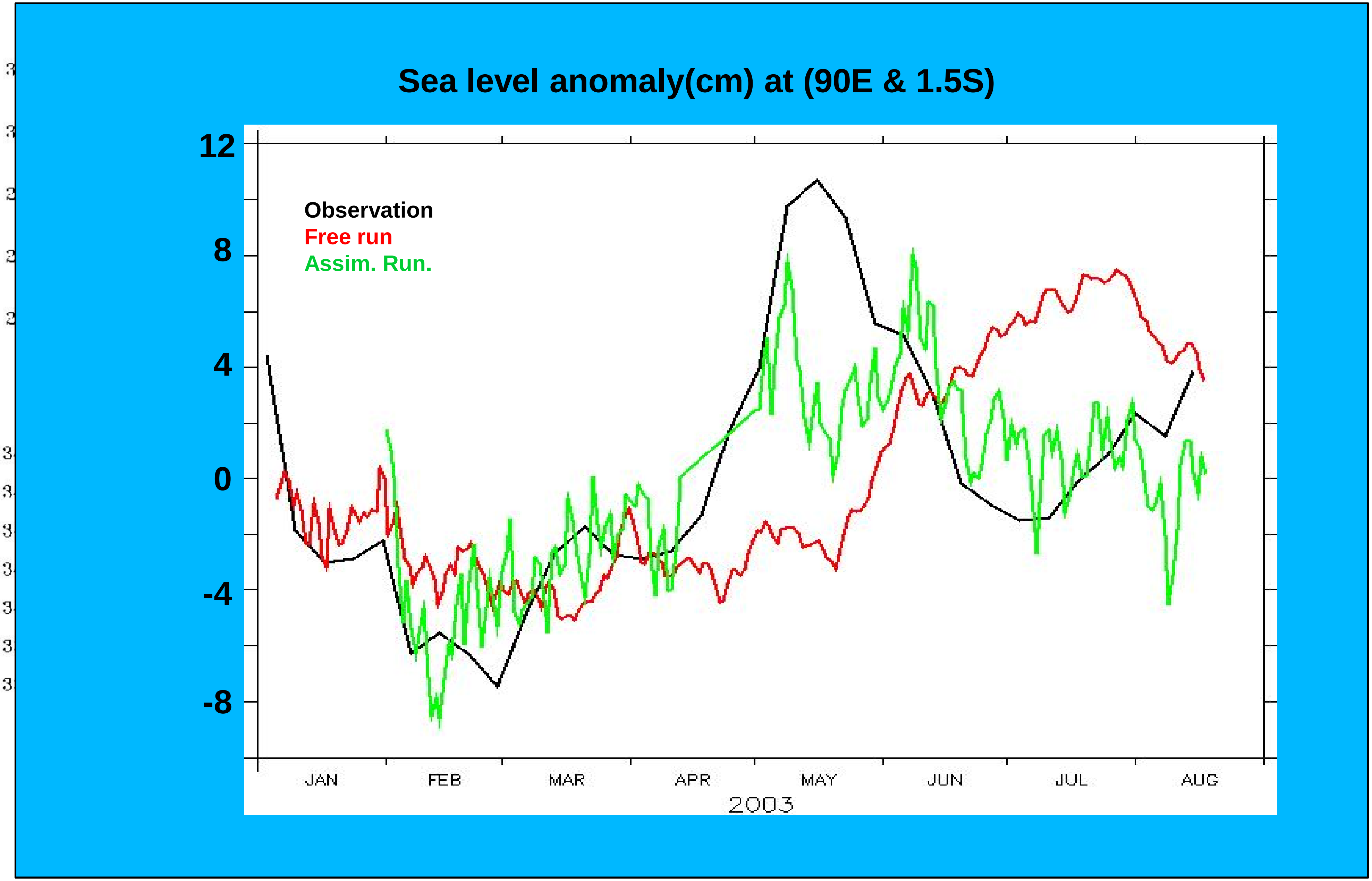
It requires more fine-scale observations to improve these schemes.

For example, it is important to understand the dynamics of the submesoscale processes (fronts) for fine-tuning the parameterisation of these processes in the models.



Map of high resolution sea surface temperature (°C) on 21 December 2010

LETKF based Data Assimilation in ROMS



- 56 Ensemble Members.
- In-situ T & S and satellite data of SST are assimilated.
- SSS & SLA not assimilated.

Evaluation of ROMS simulated surface chlorophyll in the Indian coastal waters