Fundamental Physical Oceanography

- Processes that control the surface temperature variability
- Surface forcings
- Mixed Layer processes

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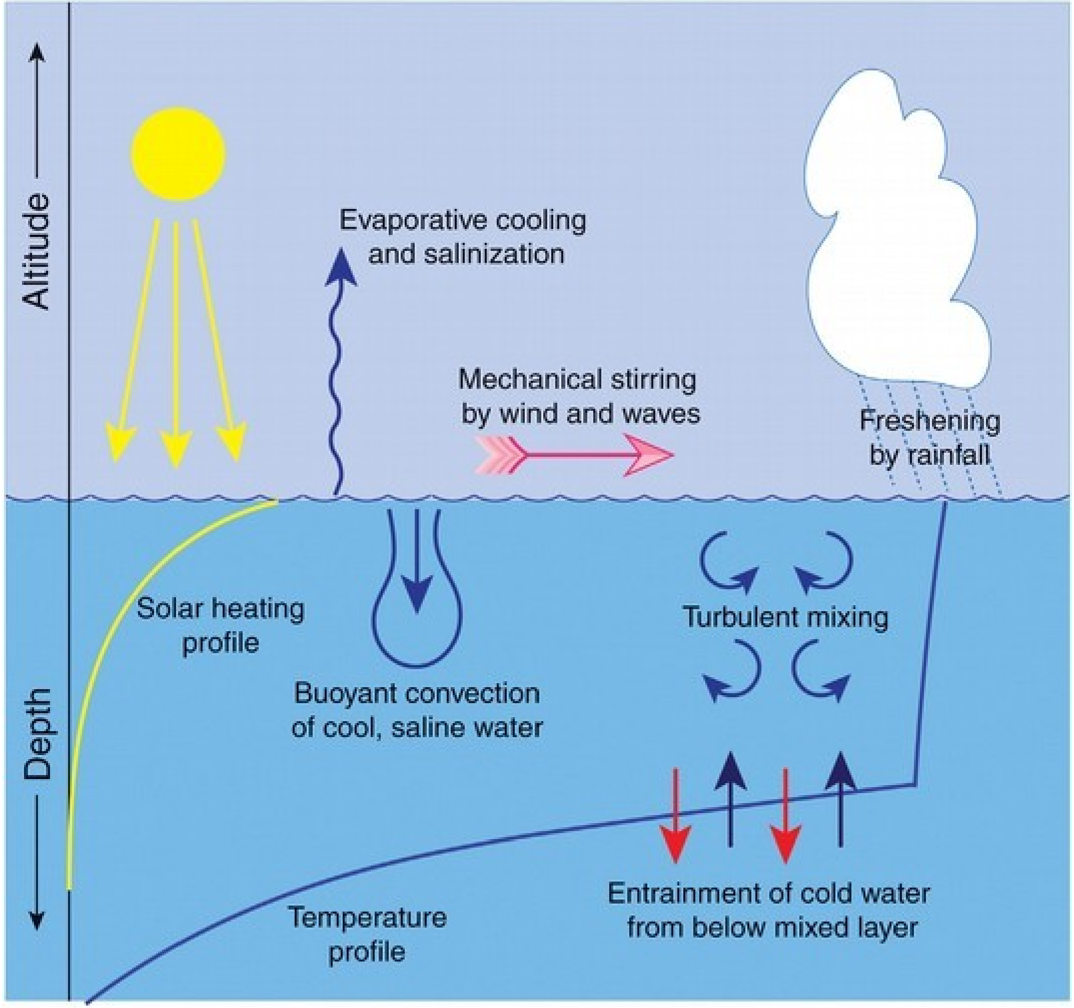
Important mixed layer processes



Global Physical Climatology, Dennis Hartmann

Processes at the air-sea interface (Model forcings)

- Incoming shortwave
- Outgoing longwave
- Latent heat flux
- Sensible heat flux
- Evaporation/precipitation
- Wind stress

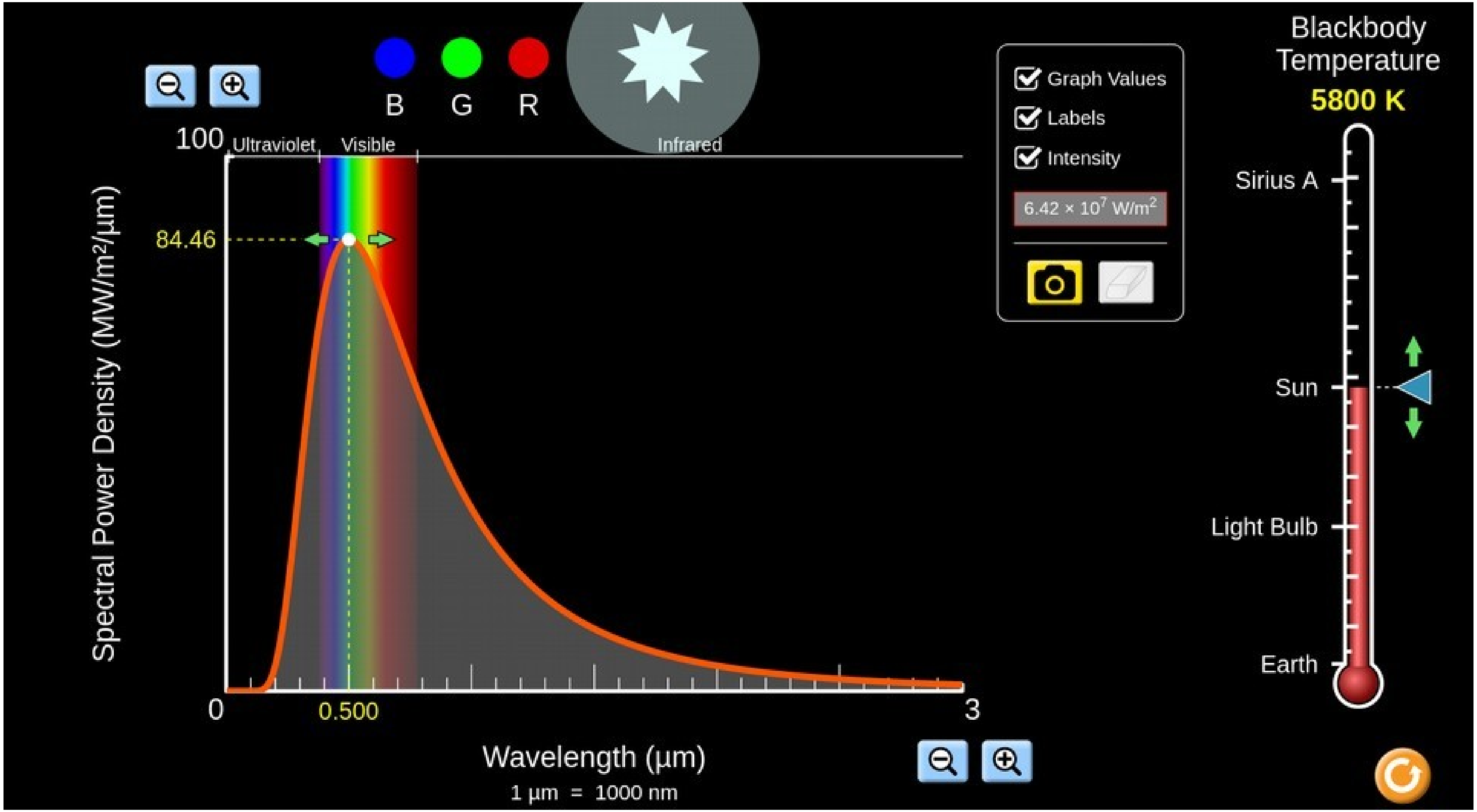
Processes at land-sea interface

- River discharge

Oceanic processes

- Vertical and horizontal mixing
- Vertical and horizontal advection
- Entrainment/detrainment
- Penetrative SW radiation

Shortwave radiation (wavelength<4000 nm)



https://phet.colorado.edu

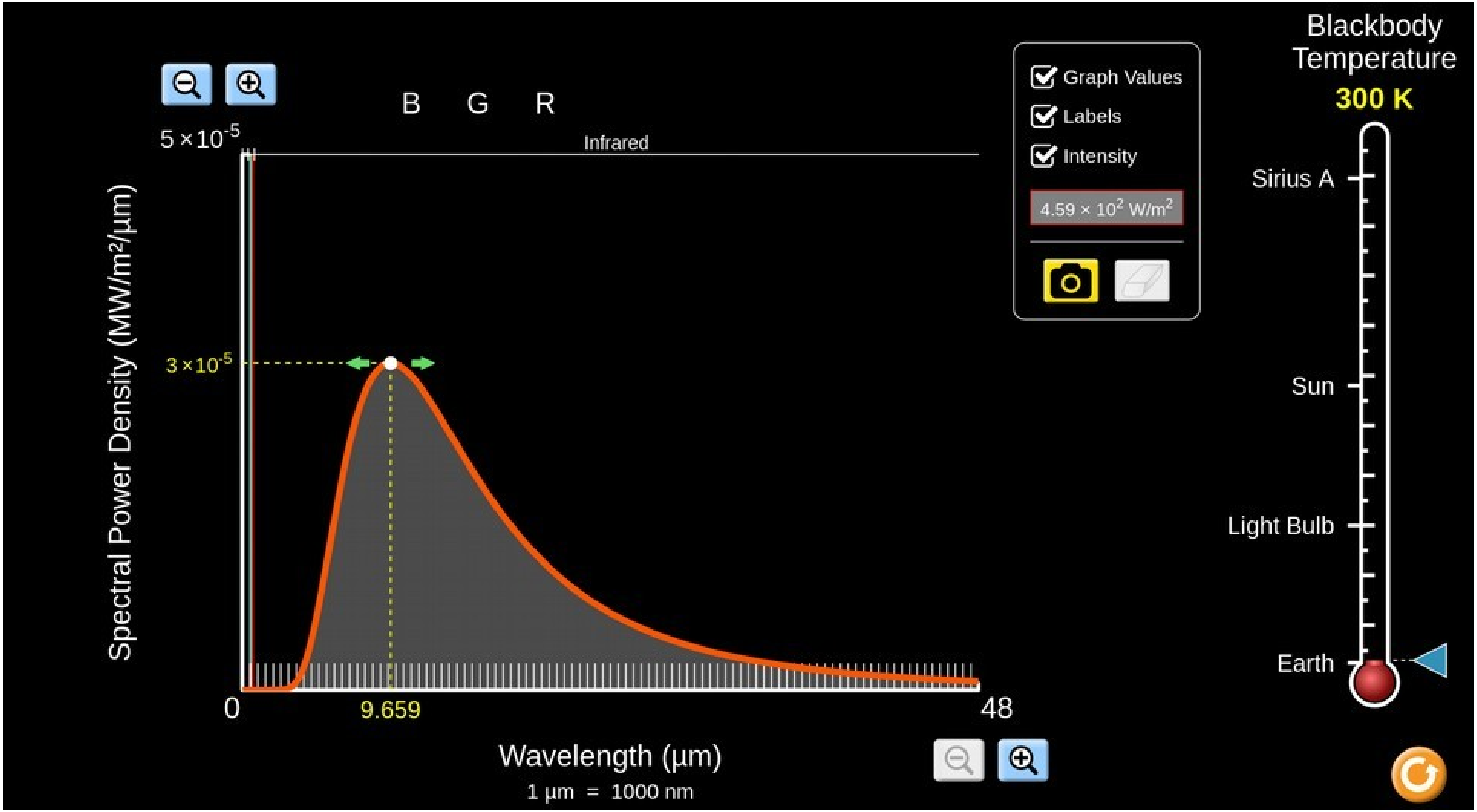
Shortwave radiation

$$Q_{SW} = S_0 \phi (1 - \alpha)(1 - 0.7 n_c)$$

- Here S_0 is the solar constant in W m^{-2}
- Φ is a function of the latitude, time of the day and season
- α is known as the albedo (a measure of reflectance)
- n_c is a measure of scattering and absorption by the sky (clouds)
- (See Gill, 1982; Talley et al., 2011; Cronin et al., 2019) for details
- Use the above equation along with satellite-derived data for clouds, water vapour content and reflectance

Photodiode Pyranometer

Outgoing longwave radiation (wavelength>4000 nm)



https://phet.colorado.edu

Longwave radiation/Back radiation

$$Q_{LW} = -\epsilon\sigma T_s^4(0.39 - 005\, e^{1/2})(1 - k\, n_c^2) - 4\,\epsilon\sigma T_S^3(T_s - T_a)$$

- Here ε is the emittance of the sea surface (0.98)
- Σ is Stefan Boltzmann constant
- T_s is the sea surface temperature in K
- T_a is the air temperature at 10 m height in K
- e is the vapour pressure at the same height
- k is the cloud cover coefficient and is a function of the latitude
- n_c is a measure of scattering and absorption by the sky (clouds)
- First term is corrected for downwelling radiation by atm.; Second term is significant in areas where air-sea temperature diff is large.
- Minus sign indicate that it is a loss term
- (See Josey et al., 1999; Talley et al., 2011; Cronin et al., 2019) for details

Infrared radiometer (Pyrgeometer) – upwelling and downwelling radiation

http://www.eppleylab.com

Flux at the air-sea interface

$$F_q = C_q \rho_a v_a \Delta q$$

- This is a simple bulk aerodynamic method
- Eddy-correlation method requires 3D measurements of temp, velocity, humidity at high resolution.
- F_q gives surface flux of variable q.
- C_q is a proportionality constant.
- ρ_a is the density of the air
- v_a is the wind speed measured at 10 m
- Δq is the difference in the property between the atm (q_a) and the ocean (q_0)

Wind stress

$$\tau_x = C_m \rho_a v_a (\vec{u}_a - \vec{u}_0);\ \tau_y = C_m \rho_a v_a (\vec{v}_a - \vec{v}_0)$$

- C_m is often referred to as the drag coefficient ($\sim 10^{-3}$) and is a function of sea state, wind speed and rainfall
- Reversing monsoon winds in the north Indian Ocean
- Wind stress is north-south during the summer monsoon and westerlies during the intermonsoon months.

Wind stress curl

$$curl\, \tau = \frac{\partial \tau_x}{\partial y} - \frac{\partial \tau_y}{\partial x}$$

- It is derived from wind stress
- Very important dynamics variable
- Regions of negative (positive) curl tend to shallow (deepen) isopycnals, a process known as Ekman pumping

Evaporation

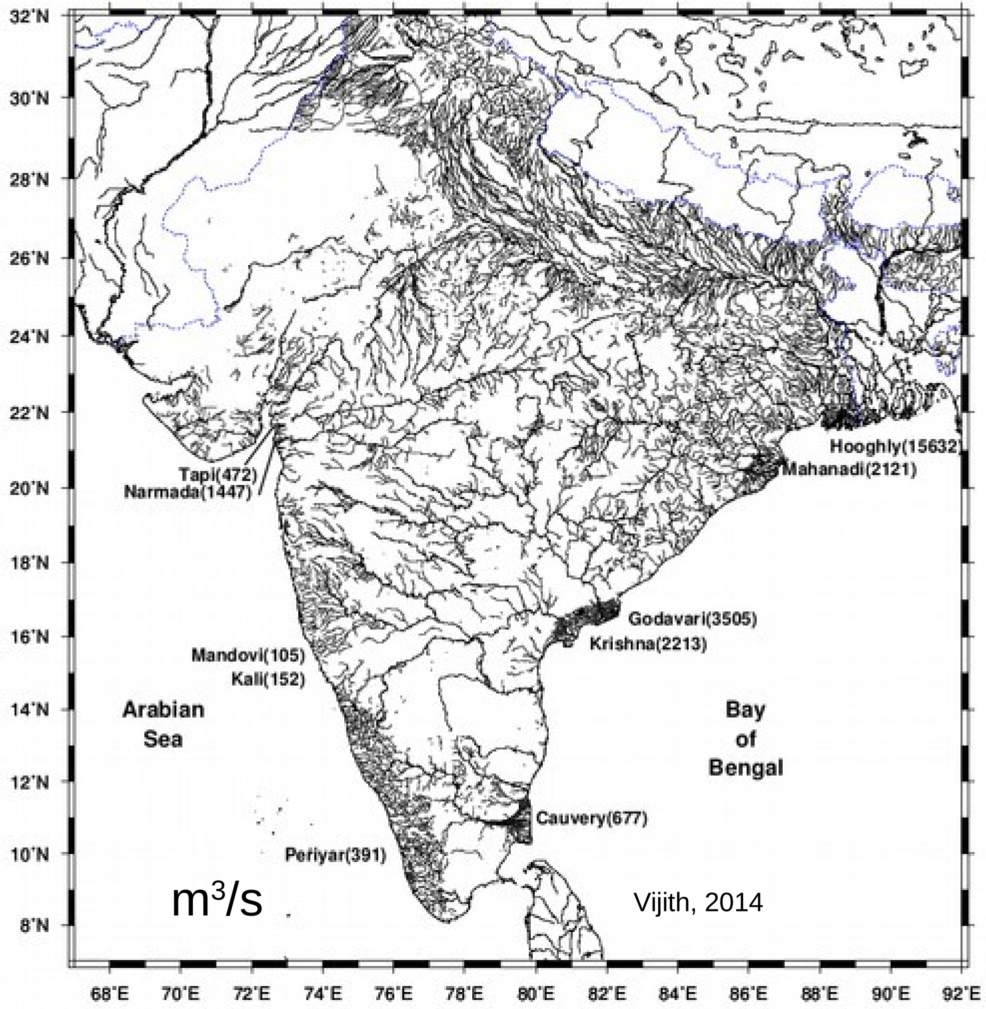
$$\epsilon = C_e \rho_a v_a (q_s - q_a)$$

- C_e is ~1.6 x 10^{-3}
- q_s saturation specific humidity at temperature Ts and q_a is specific humidity at 10 m height
- Evaporation is a diffusive process
- ε is a loss term (freshwater is lost from sea surface)

Precipitation

- Adds freshwater to ocean surface (mm/day)
- Rain gauges or satellites (OLR/microwave)

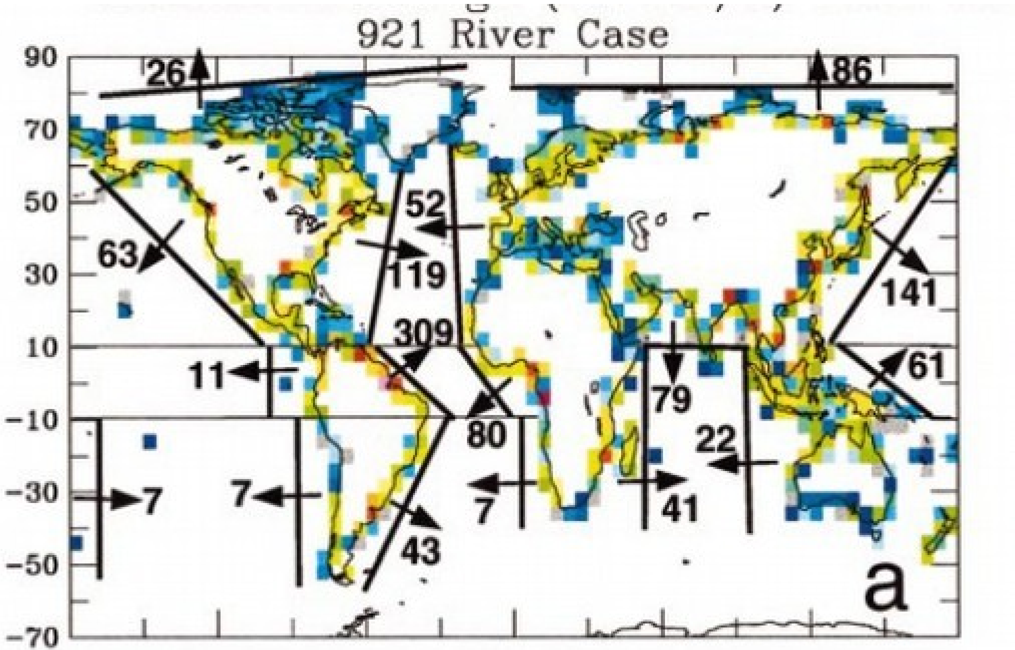
River discharge



m³/s

Vijith, 2014

1000 m³/s



Dai and Trenberth, 2002

Latent Heat Flux

$$Q_L = -L\epsilon = -LC_e\rho_a v_a(q_s - q_a)$$

- C_e is ~1.6×10^{-3}
- L is latent heat of evaporation 2.44×10^6 J/Kg.
- The minus sign indicate that the ocean is cooled by evaporation

Sensible Heat Flux

$$Q_S = -C_S\rho_a v_a C_P(T_s - T_a)$$

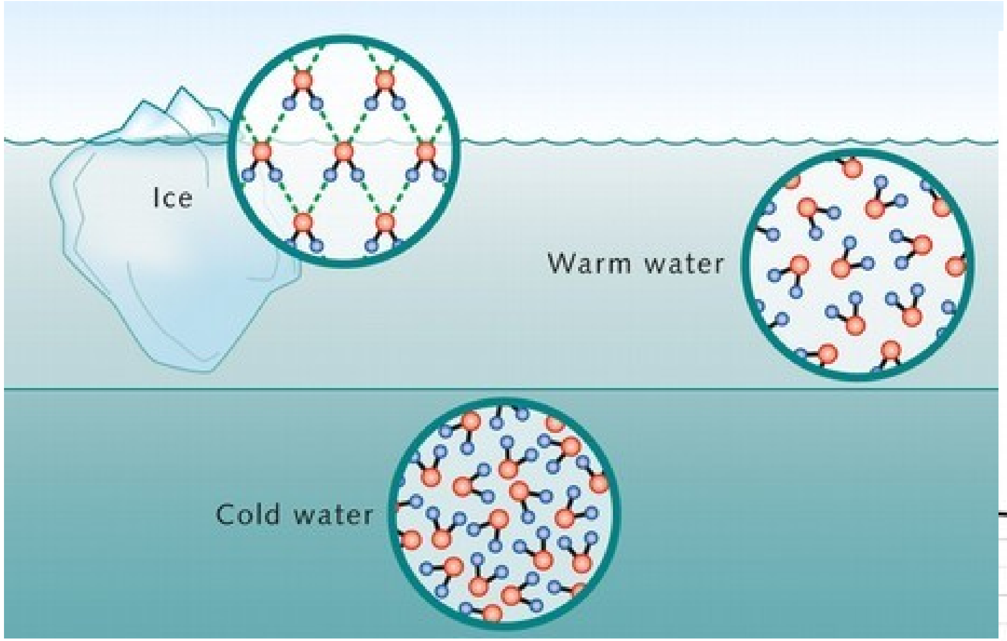
- C_S is ~1.6×10^{-3}
- C_P is specific heat capacity of air 1004 J/Kg/^{o}C.
- Q_S is usually negative and cools the ocean

Net heat flux

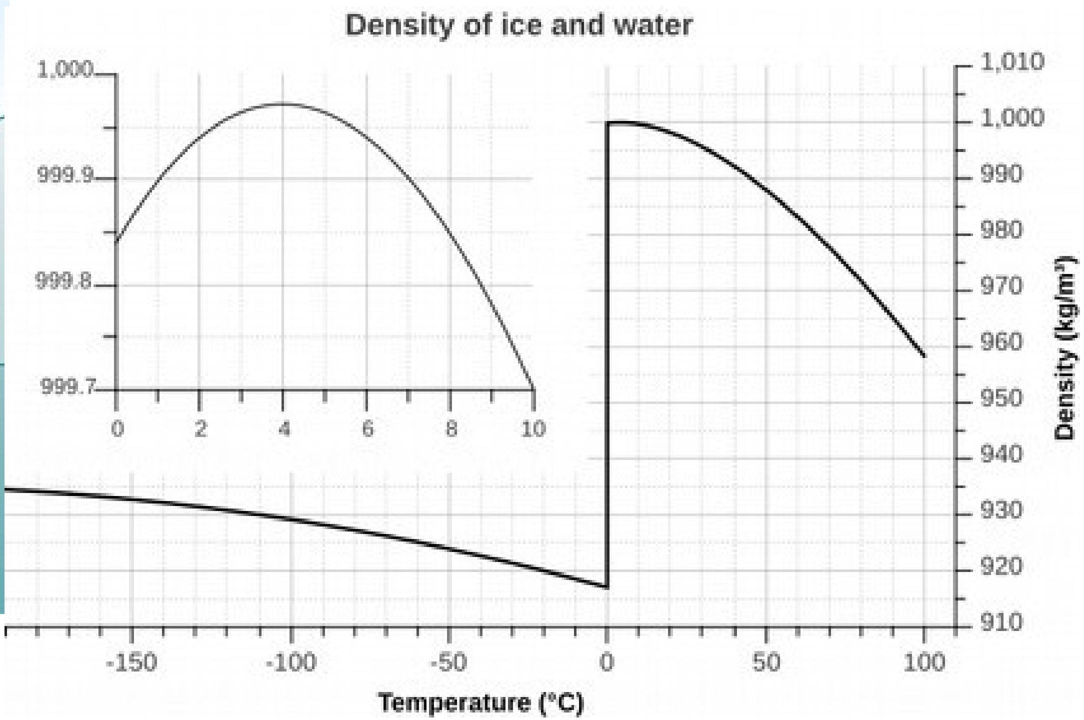
$$Q = Q_{SW} + Q_{LW} + Q_L + Q_S$$

- Q shows a complex pattern of variability
- For the north Indian Ocean Q is positive and is balanced by a cross-equatorial meridional overturning circulation

Density of water as a function of temperature

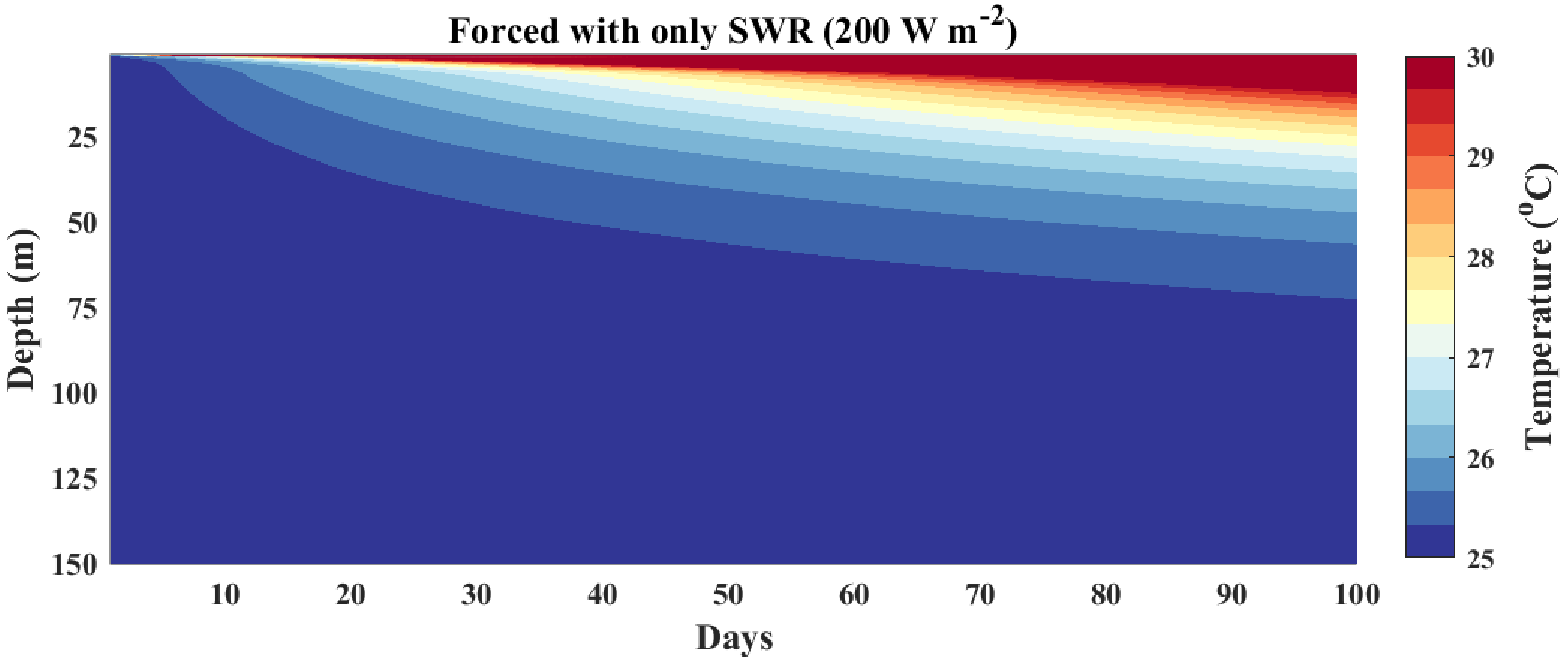


https://worldoceanreview.com

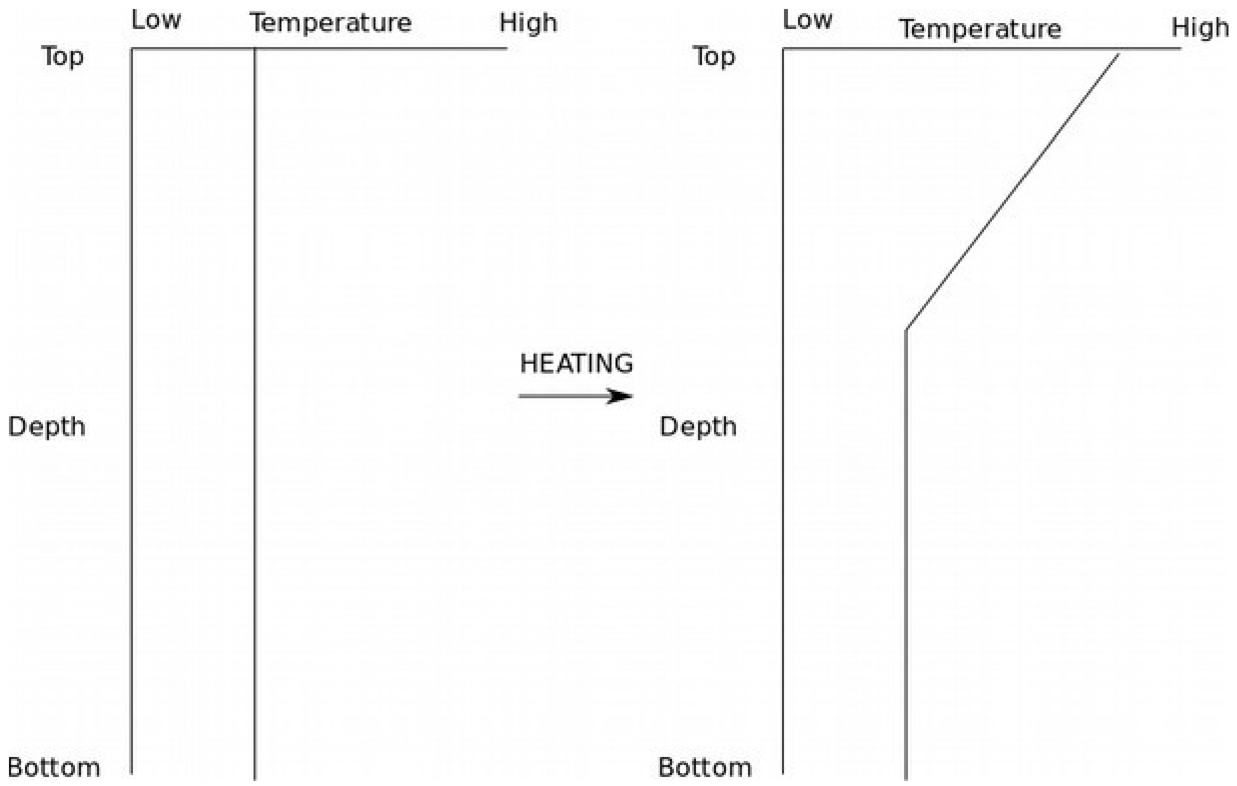


https://en.wikipedia.org

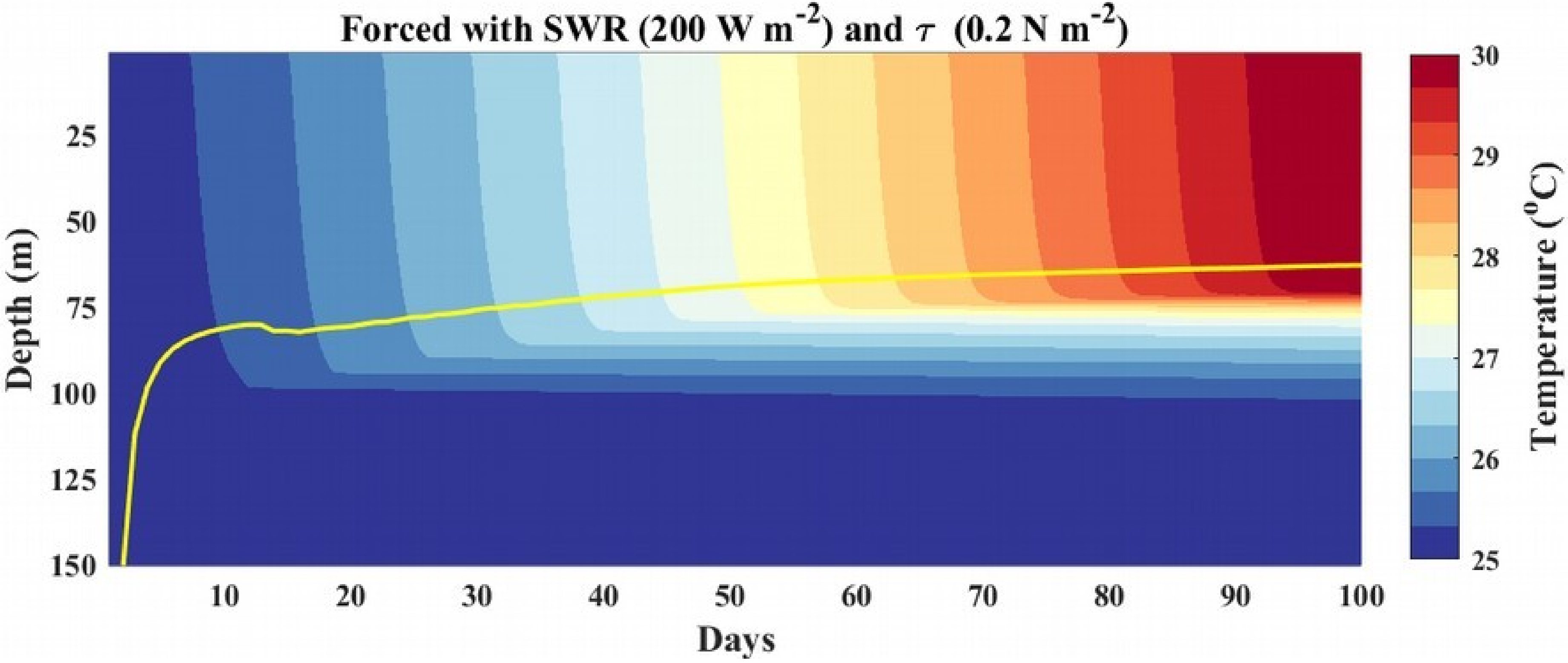
Heating from the top



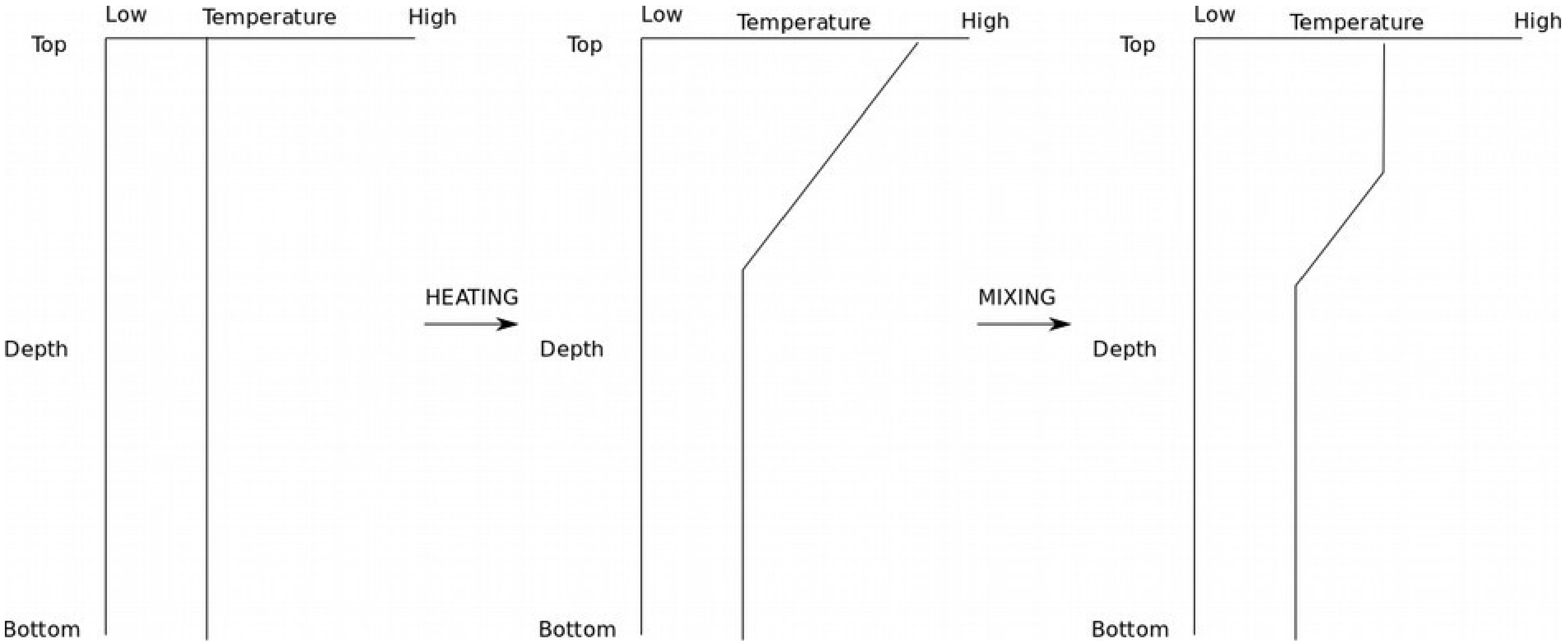
Heating from the top



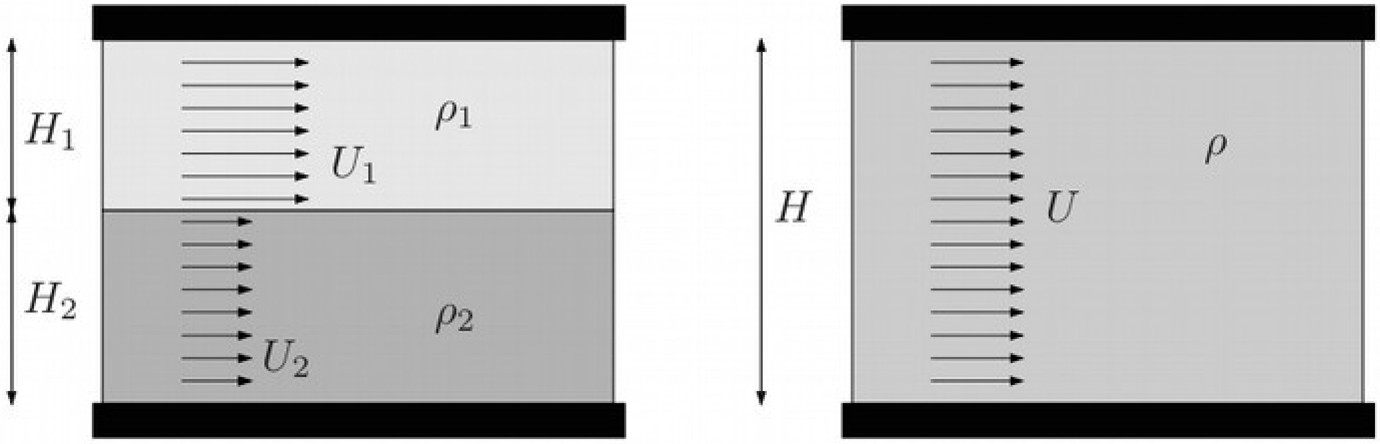
Heating from the top + mixing at the surface



Heating from the top + mixing at the surface



Mixing of stratified fluids

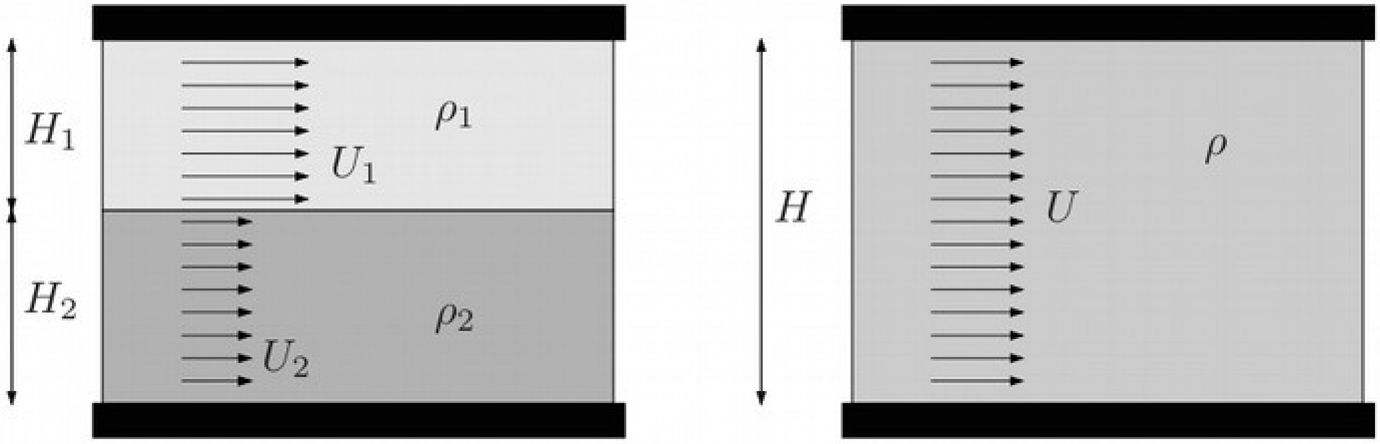


Cushman-Roisin and Beckers, 2009

- Assuming H_1=H_2=H/2; ρ=(ρ_1+ρ_2)/2

$$PE\,gain = \int_0^H \rho_{final}\, gz\, dz - \int_0^{H/2} \rho_2\, gz\, dz - \int_{H/2}^{H} \rho_1\, gz\, dz = \frac{1}{8}(\rho_2 - \rho_1)\, gH^2$$

Mixing of stratified fluids



Cushman-Roisin and Beckers, 2009

- Assuming $\rho_1=\rho_2=\rho_0$; $u=(u_1+u_2)/2$

$$KE\ loss=\int_0^{H/2}\frac{1}{2}\rho_0 u_2^2\,dz+\int_{H/2}^{H}\frac{1}{2}\rho_0 u_1^2\,dz-\int_0^{H}\frac{1}{2}\rho_0 u_{initial}^2\,dz=\frac{1}{8}\rho_0\left(u_1-u_2\right)^2 H$$

Mixing of stratified fluids

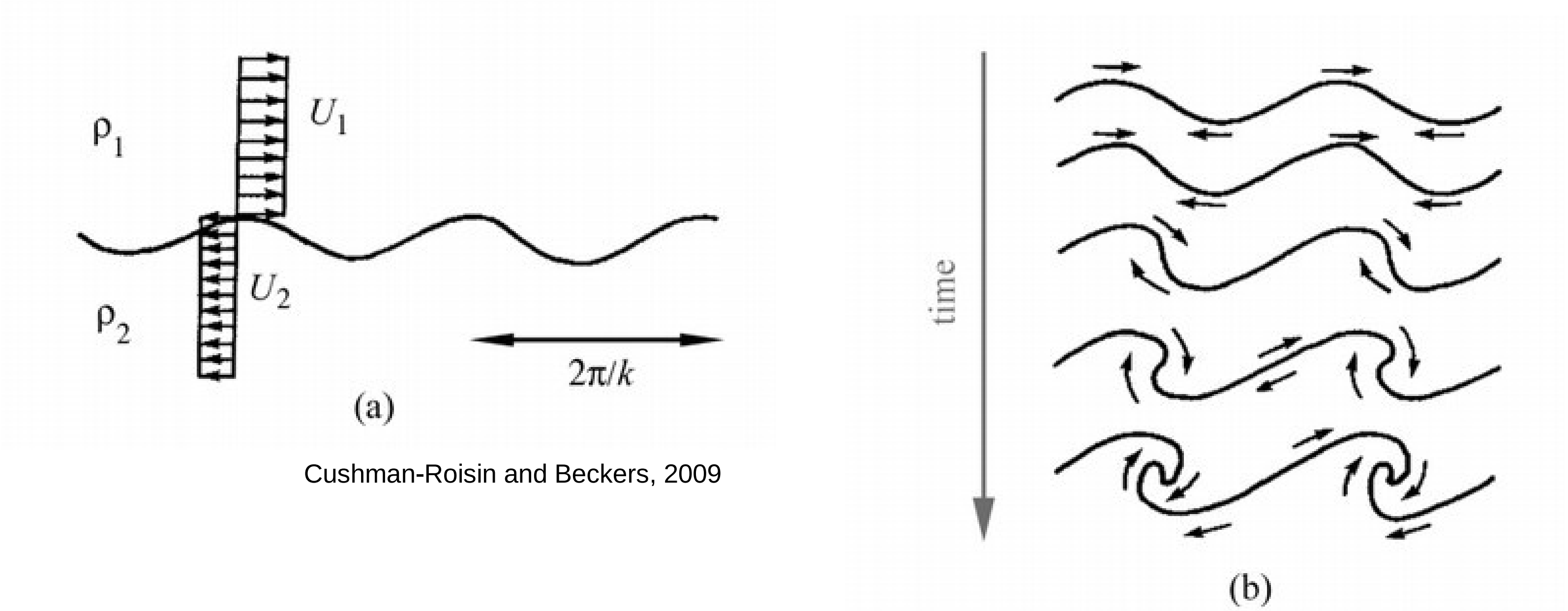
- Mixing happens when KE loss ≥ PE gain

$$\frac{1}{8}\rho_0(u_1-u_2)^2 H \geq \frac{1}{8}(\rho_2-\rho_1) g H^2$$

$$\rho_0(u_1-u_2)^2 \geq (\rho_2-\rho_1) g H$$

- When the inequality is satisfied, the interface becomes unstable due to Kelvin-Helmholtz instability developing small-amplitude undulations; they rapidly grow to form large-amplitude waves that eventually break.

Kelvin-Helmoholtz instability

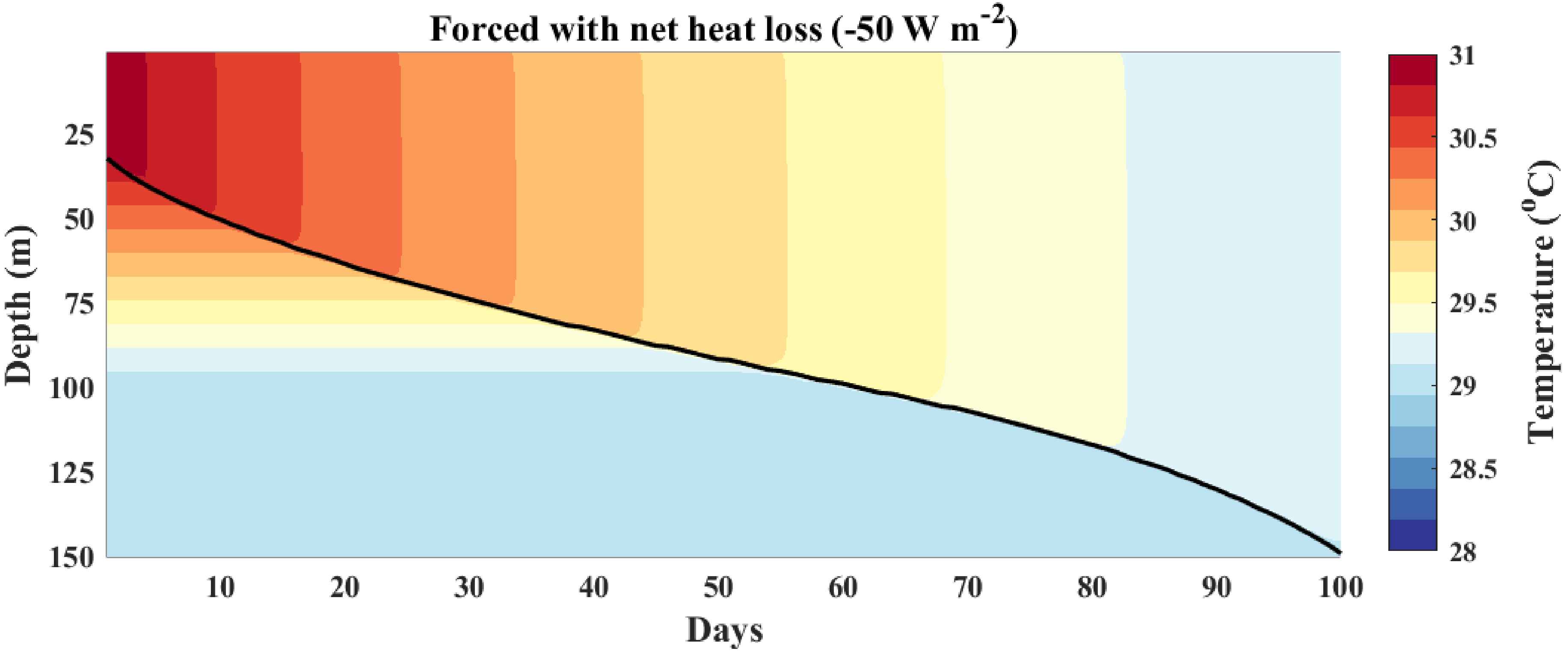


Cushman-Roisin and Beckers, 2009

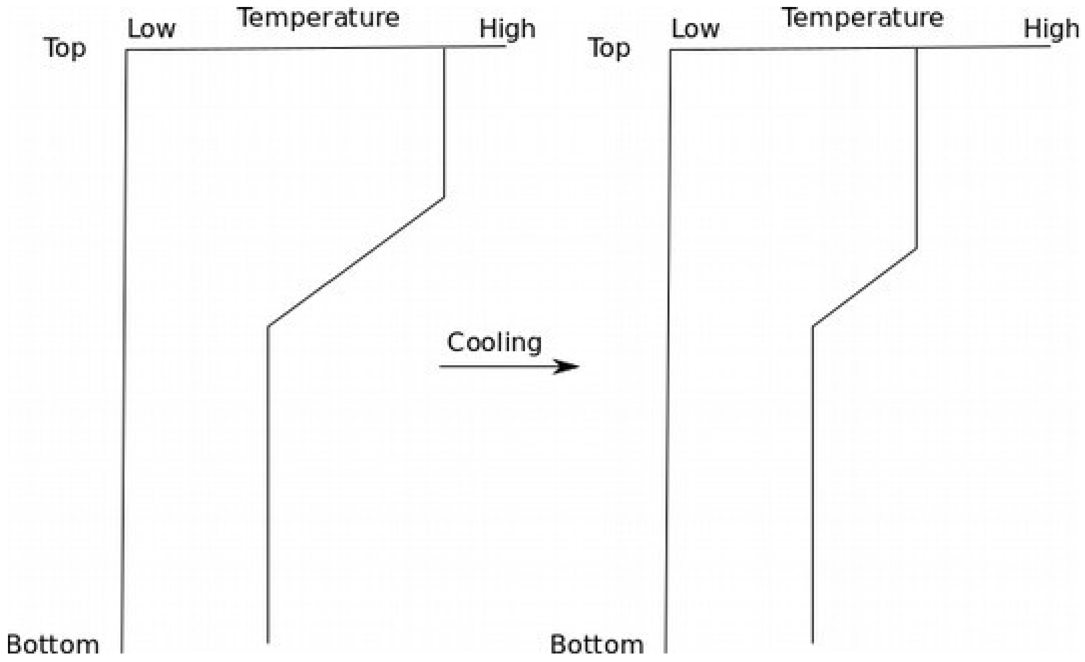
- A more rigorous analysis of the two layer system of infinite extent is available in Kundu and Cohen, 2002
- The instability criteria derived for such a system is known as Kelvin-Helmoholtz instability

$$\rho_1 \rho_2 k (u_1 - u_2)^2 \geq (\rho_2^2 - \rho_1^2) g$$

Heat loss at the surface



Heat loss at the surface



Mixing by cooling/Buoyancy flux

- Surface density of the water can be increased by a negative Q or (P-E)

$$\beta = \frac{Dg}{\rho_0} = -\alpha_t \frac{Q}{C_p} - \rho_0 \alpha_s (P - E) S_m$$

- Here, α_t and α_S are coefficients of thermal expansion and saline contraction, respectively
- D is density flux
- β is buoyancy flux
- S_m is salinity

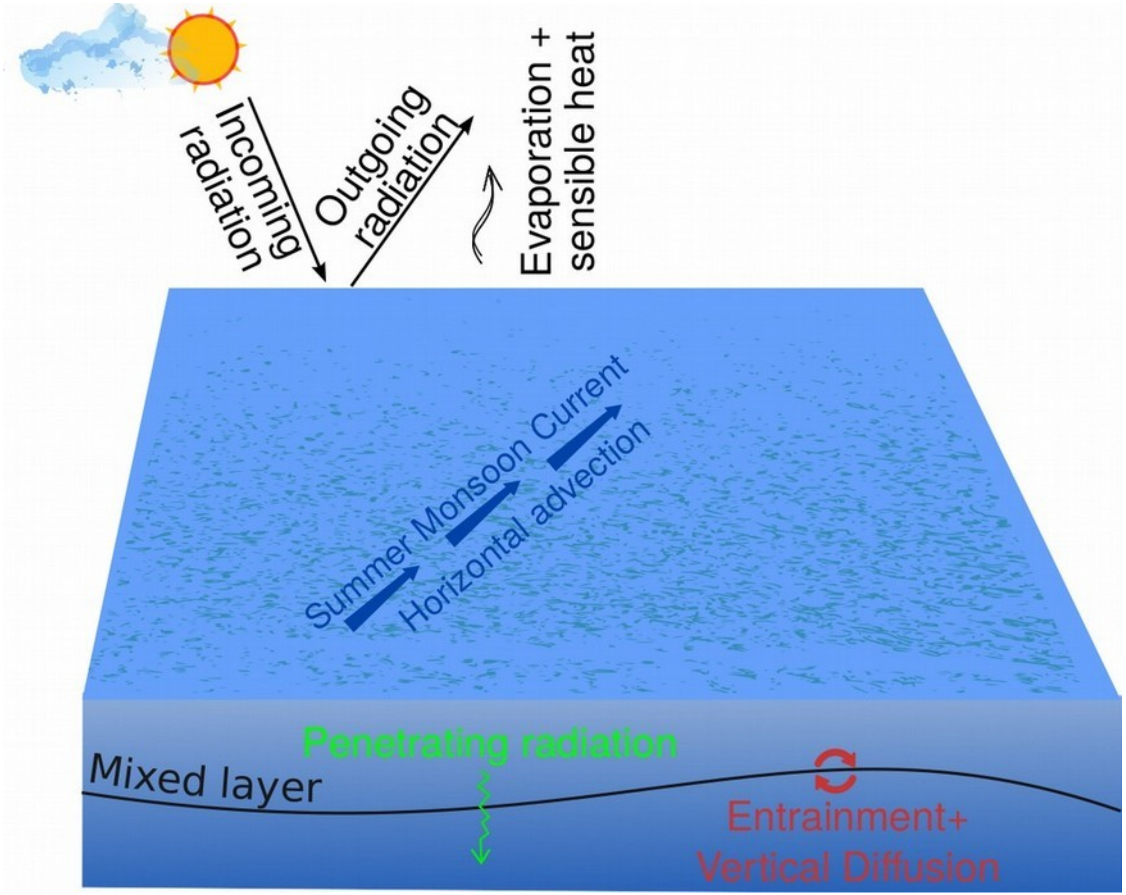
Rayleigh-Benard Instability

- The first intensive experiments on instability caused by heating a layer of fluid were conducted by Benard in 1900 (Kundu and Cohen, 2002). Benard observed beautiful hexagonal cells when the convection began. Later experiments conducted in thicker cells showed convection cells of many shapes (not only hexagonal).
- Raleigh in 1916 proposed that instability would occur when the adverse vertical gradient of temperature is large enough so that the following ratio (Rayleigh number) exceeded a critical value.

$$Ra = \frac{g \alpha \Gamma h^4}{\kappa \nu}$$

- α is coefficient of thermal expansion
- Γ is -dt/dz, vertical temperature gradient
- h is the layer depth
- κ is thermal diffusivity
- ν is kinematic viscosity
- It can be shown that the wavelength at the onset of the instability is 2h.

Heat storage in the ML



Vijith et al., 2020

Heat storage in the ML

$$\underbrace{\frac{\partial T_a}{\partial t}}_{\text{Tendency}} = \underbrace{-\left(u_a\frac{\partial T_a}{\partial x} + v_a\frac{\partial T_a}{\partial y}\right)}_{\text{Horizontal advection}} + \underbrace{\kappa_H\left(\frac{\partial^2 T_a}{\partial x^2} + \frac{\partial^2 T_a}{\partial y^2}\right)}_{\text{Horizontal mixing}} - \underbrace{\frac{1}{h}\left[\kappa_Z\frac{\partial T}{\partial z}\right]_{-h}}_{\text{Vertical mixing}}$$

$$\underbrace{-\left(\frac{T_a - T_{-h}}{h}\right)\left(\underbrace{\frac{\partial h}{\partial t}}_{\text{ML tendency}} + \underbrace{u_{-h}\frac{\partial h}{\partial x} + v_{-h}\frac{\partial h}{\partial y}}_{\text{Lateral induction}} + \underbrace{w_{-h}}_{\text{Vertical advection}}\right)}_{\text{Entrainment}}$$

$$+ \underbrace{\frac{q_0 - q_{pen}}{\rho_0 c_p h}}_{\text{Net heat flux}}.$$

Vijith et al., 2020

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The Seasonal Heat Budget of the North Pacific: Net Heat Flux and Heat Storage Rates (1950–1990)

JOHN R. MOISAN* AND PEARN P. NIILER

Physical Oceanography Research Division, Scripps Institution of Oceanography, La Jolla, California

(Manuscript received 19 March 1996, in final form 15 July 1997)

Heat Conservation Equation

a. Heat storage rate

Because of the role that vertical motion plays in the upper-ocean heat budget (Emery 1976), a heat storage rate equation was chosen that calculated the amount of heat stored down to a chosen isotherm. Here is a formal derivation of this heat storage rate equation.

We begin with the conservation of mass equation,

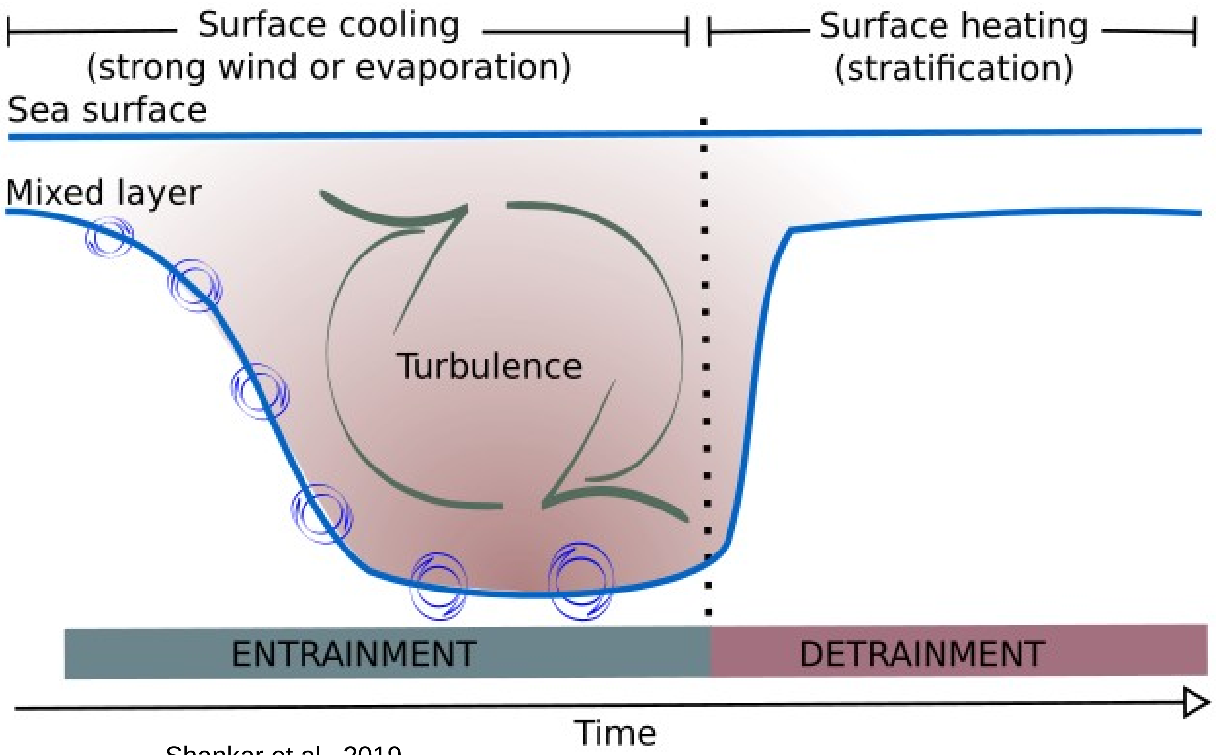
$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \qquad \text{(A1)}$$

and the conservation of heat equation,

$$\rho c_p \left[\frac{\partial T}{\partial t} + u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y} + w\frac{\partial T}{\partial z} \right] = \frac{\partial q}{\partial z}, \qquad \text{(A2)}$$

where ρ, c_p, and T are the mean density, specific heat, and temperature of seawater, respectively, and q is the vertical heat flux. By multiplying Eq. (A1) by T and dividing Eq. (A2) by ρc_p, we can add both equations to get

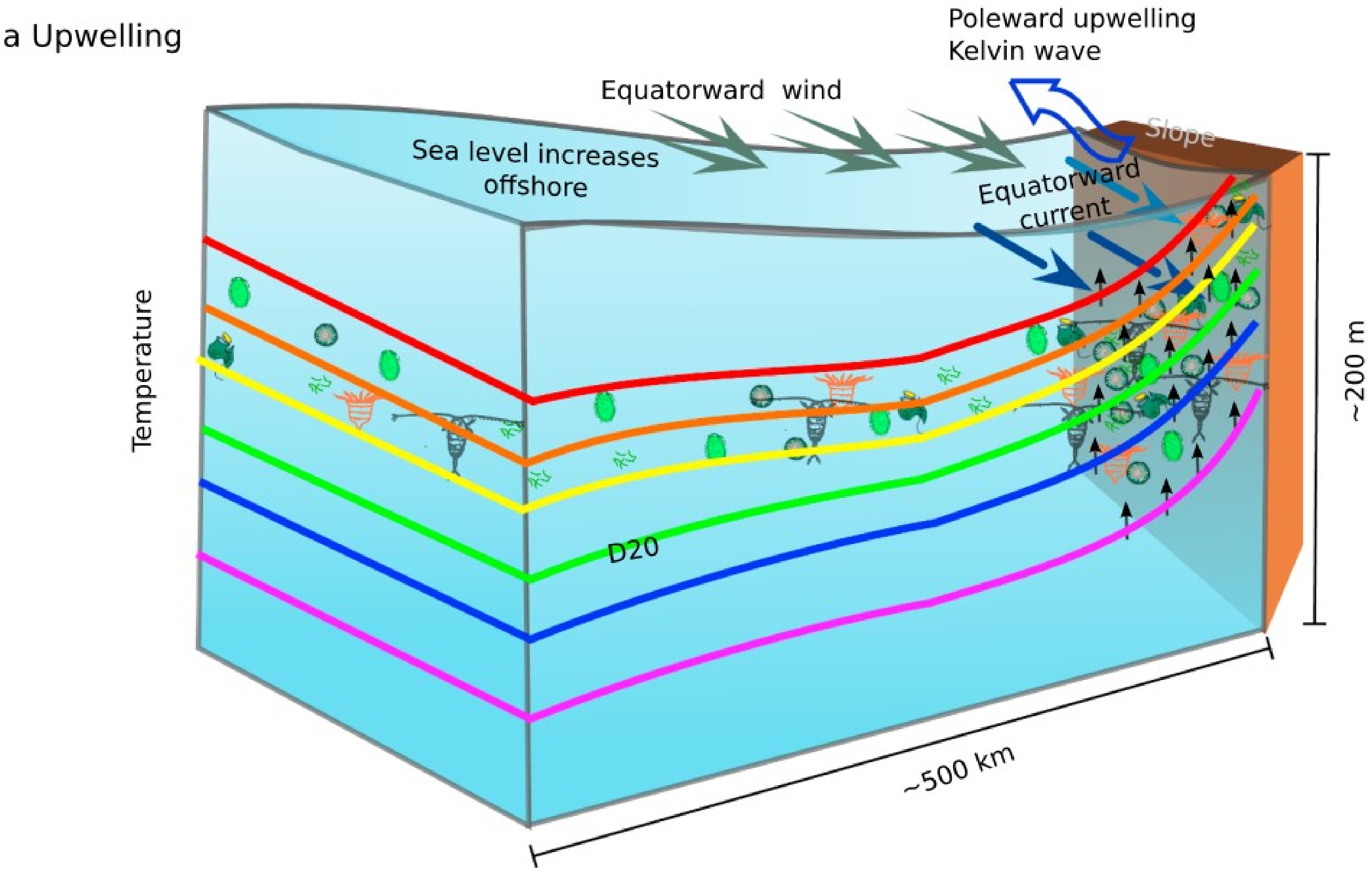
Entrainment/detrainment



Shankar et al., 2019

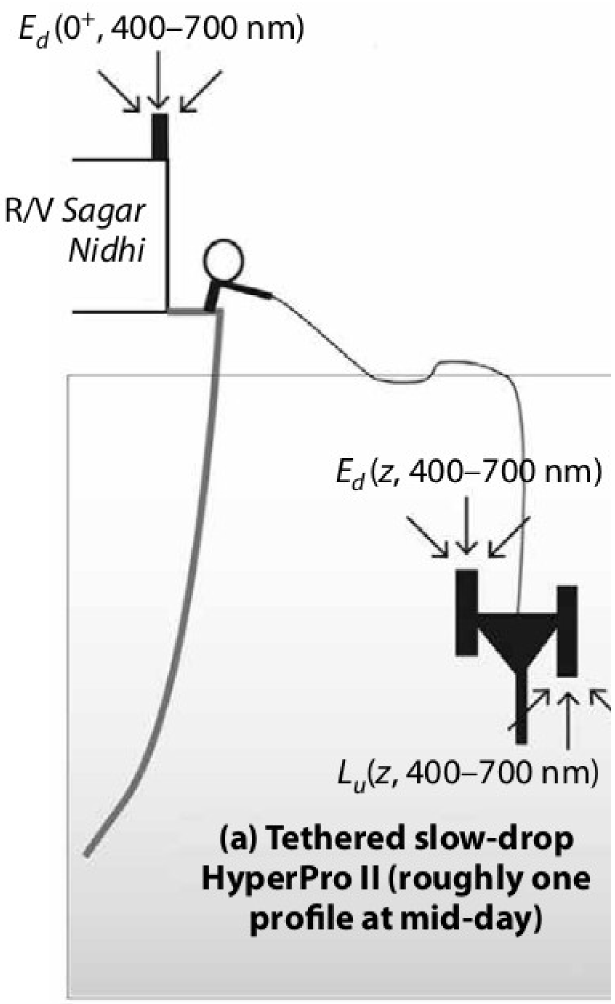
ML turbulence can still mix water from just below the ML into the layer so that ML gradually thickens at the rate dh/dt, a process known as "entrainment."

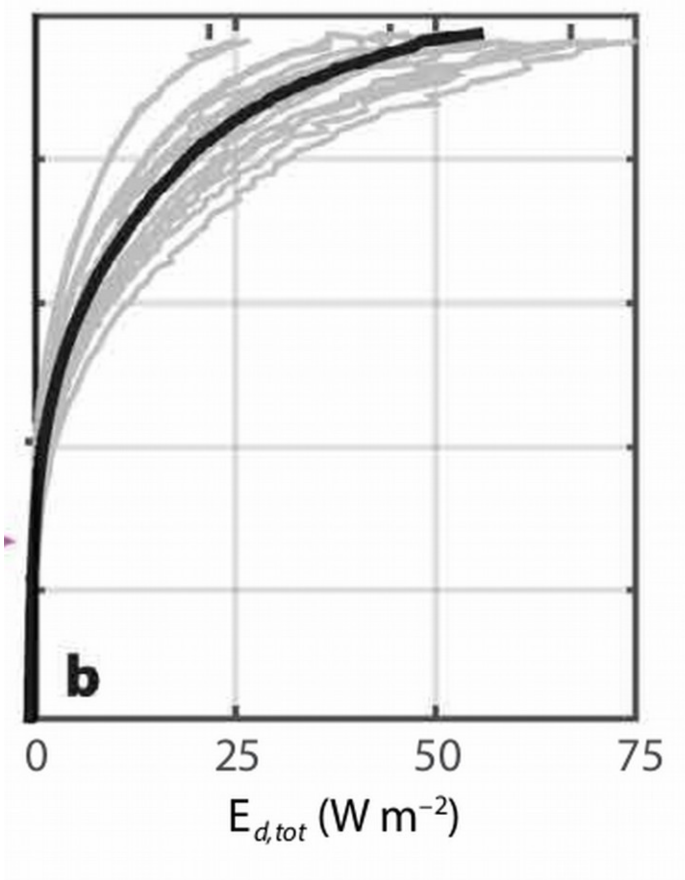
Vertical advection or upwelling



Shankar et al., 2019

Penetrating shortwave radiation

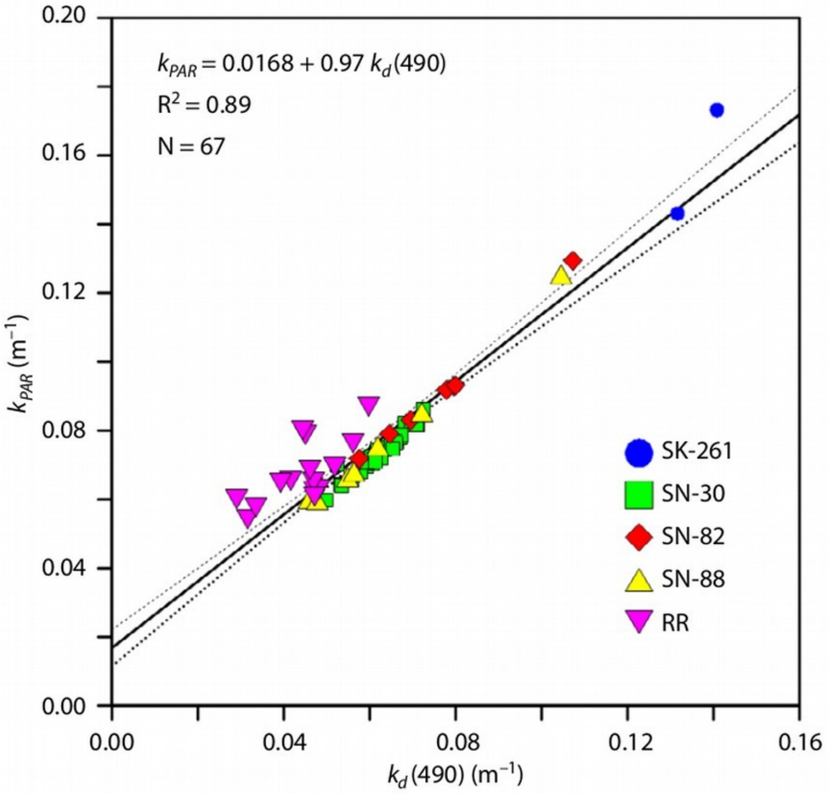




Lotliker et al., 2016

Penetrating shortwave radiation

$$Q_{pen} = Q_0 e^{k_{PAR} h}$$



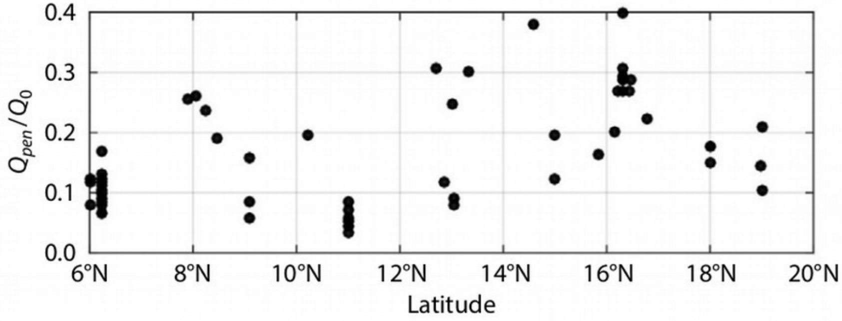
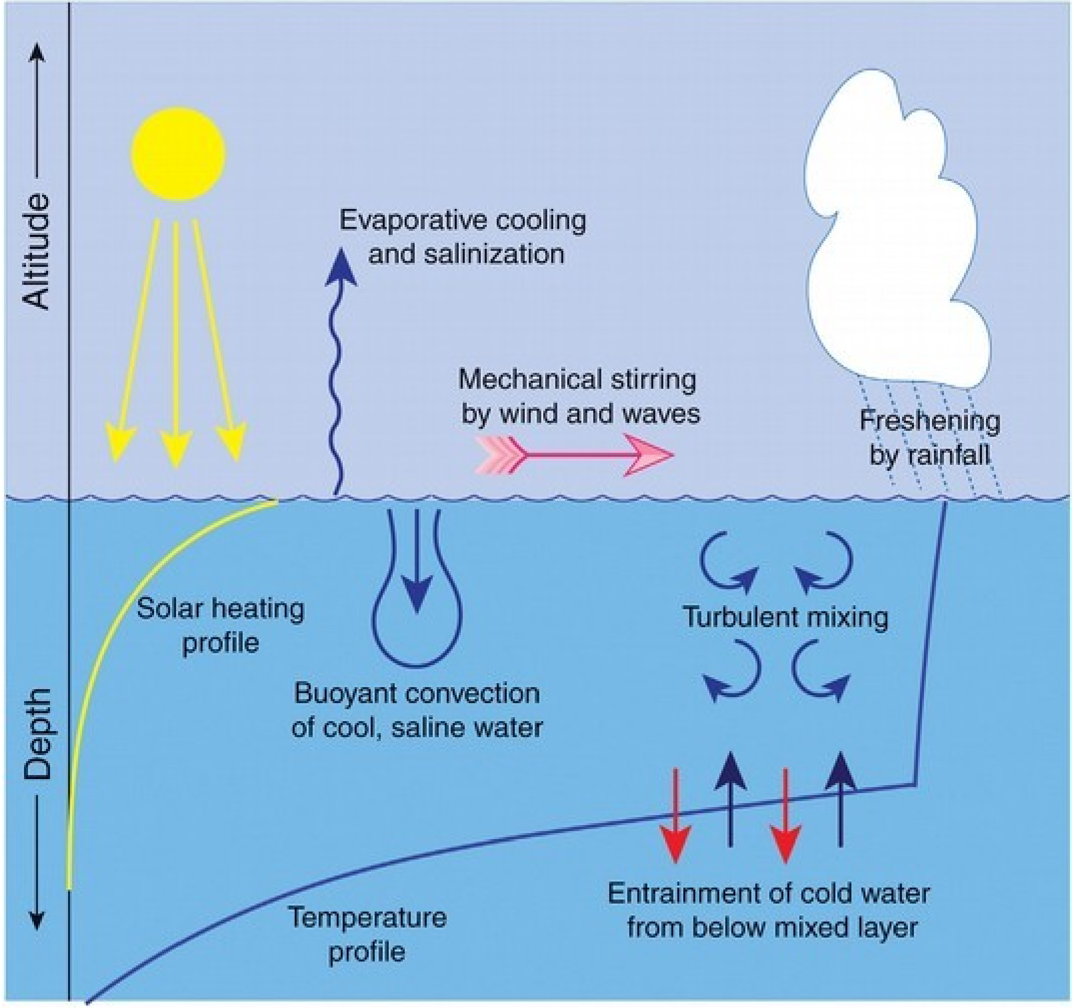


FIGURE 6. Latitudinal variability of the ratio of heat that penetrates below the mixed layer depth relative to the incident shortwave heat flux.

FIGURE 5. Scatter plot showing the relation between the downwelling diffuse attenuation coefficient at 490 nm $k_d(490)$ and diffuse attenuation coefficient of photosynthetically available radiation (k_{PAR}). The symbols represent the sampling during various cruises as described in Figure 1. The solid line represents the trend and the dotted lines are at 95% confidence level.

Lotliker et al., 2016

Important mixed layer processes



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